ИЗВЕСТИЯ
АКАДЕМИИ НАУК СССР
СЕРИЯ ГЕОЛОГИЧЕСКАЯ

IZVESTIYA AKAD. NAUK SSSR
SERIYA GEOLOGICHESKAYA

)61 CONTENTS No. 12, December

BRIEF COMMUNICATIONS

TWENTY-SIXTH YEAR OF PUBLICATION

T-11879. Approved for printing November 20, 1961.
Circulation - 3.240 copies. Order 4016

Paper size 70 x 108-1/16 | Paper 4 | Printing sheets 10.96 | Publ. sheets 12.7

Second printing office of the USSR Academy of Sciences Publishing House.
Moscow, Shubinskiy per. 10.

'HE ORIGIN OF SPHERULITIC LAVAS IN SOUTH SIKHOTE-ALIN[1]

by

I. Z. Bur'yanova, and M. A. Favorskaya

After many years of studying the Tertiary usives of South Sikhote-Alin, I. Z. Bur'yanova s established a number of locations in them ich contain occurrences of peculiar ball-like l spherulitic acid lavas. They are especially undant in certain Paleogene formations where y occur throughout the section. For innce, spherulitic lavas in the Notto River sin have been observed in the lower section the so-called Bogopol'sk formation (Paleone), while in the Imanka River basin they occ r in the upper deposits of the Brusilovsk igocene formation. Locally, they occur roughout the entire sequence, alternate with ier acidic effusives, and cover large areas.

The marked areal persistency of the "lava lls" makes them a good local datum horizon correlating individual effusive members and rmations; for this reason, they should be ven due consideration in geologic mapping. ieir main interest to us, however, is in conction with the liquation of magmas.

Liquation phenomena in the lavas of various gions have been described by F. Yu. Levinsonssing [7, 8], D. S. Belyankin [1], I. M. lovikova [3], V. I. Lebedinskiy and Mo -Min' [6], O. P. Yeliseyeva [5], and others.

The feasibility of liquation of a silicate melt s been demonstrated experimentally by D. P. igor'yev [4].

The structural features and composition of e Tertiary lava-ball and spherulitic acid vas from Sikhote-Alin suggest that they iginated during the liquation of a magmatic elt. Given below is a description of these vas from the Brusilovsk Oligocene formation posed on the upper reaches of the Imanka ver (the Bol'shaya Sinancha basin), on the est slope of Sikhote-Alin.

The following three Paleogene volcanic formations (from bottom to top, Figure 1) occur in this section:

1) the Samarga formation, presumably Paleocene, represented by alternating andesites, andesite porphyrites, less commonly by andesite-basalts, dacites, and rare intercalations of porphyrite tuffs. It rests unconformably on Upper Cretaceous and older formations. Its thickness is 800 m.

2) the Bogopol'sk formation unconformably deposited on the Samarga, consists of light-colored motley rhyolites and rhyolite-porphyries with subordinate felsites, tuffs, and tuffolavas, assigned to the Upper Paleocene. It is 900 meters thick.

3) the Brusilovsk formation is widely developed and is the highest in the section; it is unconformable on the Samarga and Belopol'sk formations and is supposed to be Oligocene.

The Brusilovsk formation is subdivided into two parts: the lower 250 m are lava breccias, acid tuffs, and subordinate felsites; the upper section is mostly thin-bedded lavas, often interbedded with felsite flows containing numerous volcanic glass lenses which are either homogeneous or contain lava balls. Tuffs are almost completely missing. This upper section of the Brusilovsk formation was the subject of our study.

Vitreous lavas and felsites comprise the top of this formation. The felsites are of various light hues: pink-grey, grey, grey-green, lilac-grey, etc., often flow-banded, thin- to thick-tabular, foliated, less commonly massive and brecciated. Their joints are quartzitized, containing occasional druses of small crystals of quartz and calcite, or banded chalcedony; locally they are kaolinized and ferruginous.

Spherulitic felsites differ from these only in their texture. Some pherocrystals are large (up to 1.5 cm), often with a radial structure. Their weathered surfaces are white and nodular. Some thin-banded felsites contain bizarre, light-grey vermicular bodies, in a

O proiskhozhdenii sharovykh lav yuzhnogo hote-Alinya, (pp. 3-12).

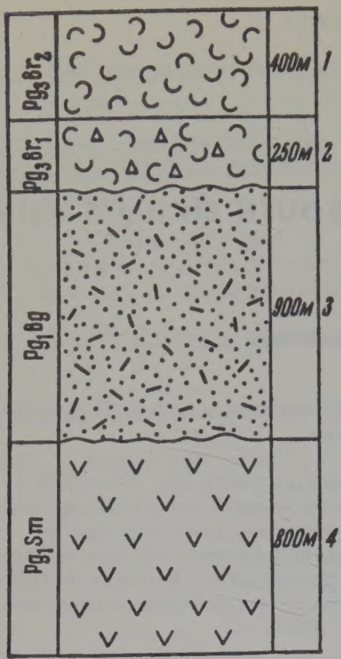


FIGURE 1. Generalized stratigraphic column of the Imanka River area effusives

1 - Upper part of the Brusilovsk formation: felsites and volcanic glass; 2 - lower part of the Brusilovsk formation: felsites and acid tuffs; 3 - Bogopol'sk formation: rhyolites, their tuffs, tuffo-lavas, rhyolite porphyries, felsites; 4 - Samarga formation: andesites, andesite porphyries, and their tuffs.

pinkish-gray groundmass (Figure 2). Each such body has an extremely fine fringe, resembling a baking crust.

The volcanic glass is quite diversified. It is represented by black (dark-gray), green, brown-green and red-brown obsidian having a vitreous luster and a conchoidal fracture; they are tabular, the tablets ranging in thickness from 0.3-0.5 cm to 2 cm, often strongly fractured, brecciated, locally containing perlite in the joints. Varieties of the lava balls are common among the black and red-brown obsidians.

Volcanic glass and varying amounts of globular formations are predominant in the lava ball beds (Figures 3 and 4). The glass is dark-grey (almost black), homogeneous, has a vitreous luster, conchoidal fracture and is cut throughout by parallel cracks and by fine, short unoriented joints. Because of the reflection from the numerous joint planes, the glass appears to be lighter in color. As seen in thin splinters, it is colorless, transparent, and contains dust-like inclusions which make it look light-grey in color. Locally, these inclusions congregate along the joints (usually in bands). Some of the shear planes are iridescent in green-blue to crimson hues. The globular masses stand out sharply against the groundmass. They are brown, usually round in cross-section, more or less consistent in size — 10 mm, seldom varying by 1-2 mm. They generally occur individually but occasionally in aggregates where their edges fuse. They are unevenly distributed, occupying from 70% to 30-40% of the rock volume. Their surface is nodular.

Under the microscope, the felsites appear to be more or less crystalline. Associated with fully crystallized . . . .
of almost . . . .
of crystallization in the above-mentioned . . .

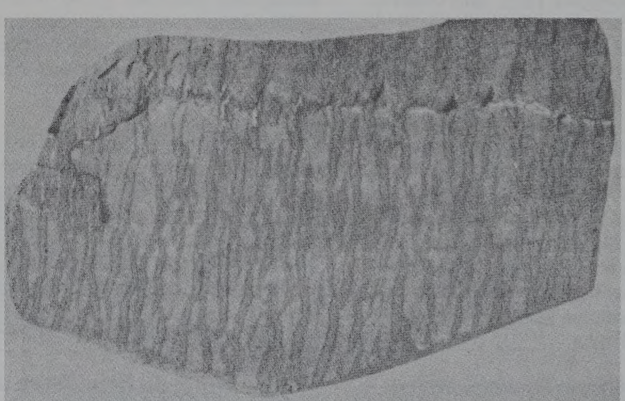


FIGURE 2. Vermicular bodies in felsites

3/4 natural size.

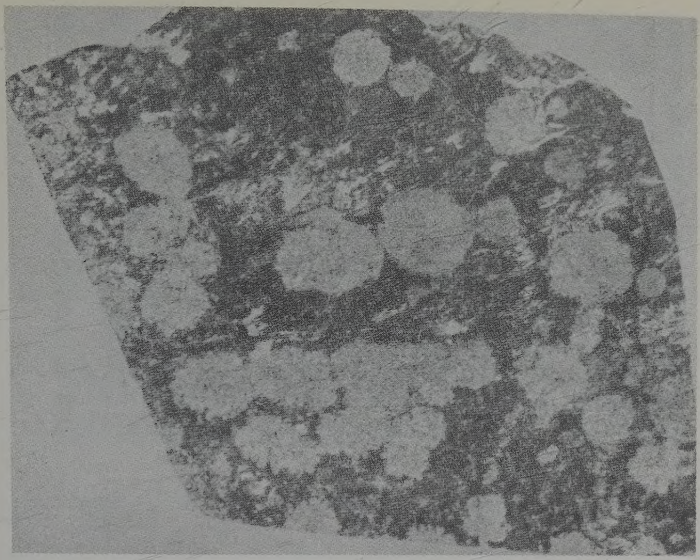
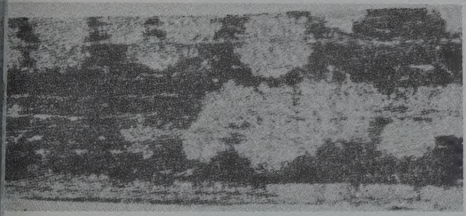


FIGURE 3. Lava balls. Section parallel to tabular parting

3/4 natural size.



GURE 4. Lava balls. Section normal to tabular parting

3/4 natural size.

nicular bodies is the same as in the en-
ing felsite.

Turning now to the lava balls it should be
tioned that their complex structure can be
n even under a binocular. They consist of
, semi-transparent, light-grey bodies with
treous luster and irregular shape — round-
oval, to palmate. They are 0.2-0.5 mm
oss, cemented with light-brown glass.
e sections show elongated lava tears ar-
ged radially, locally becoming straight and
row bands continuing beyond the balls.

The intricate structure of the lava balls
ws up particularly well in thin sections.
ee such sections were prepared — in three
ually perpendicular planes. A 5 x 8 cm
ion was cut parallel to the tabular jointing
and two 4 x 3.5 cm sections were cut in planes perpendicular to the first and to each other.

With the light under the microscope, the vitreous groundmass is colorless and includes "spheres" of dark-brown bodies immersed in a light-yellow vitreous mass. Both the cementing glass and the brown bodies contain numerous crystallites. With the Nicols crossed, the cementing glass is almost perfectly isotropic, while the brown bodies show a slight polarization in dark-grey hues, have a wavy extinction and a finely fibrous structure; the yellow vitreous mass is definitely crystallized and has a felsitic structure (Figure 5). Occasional incrustations of andesine are present in the cementing glass and in the spherules.

The three mutually-perpendicular sections show that the shape of the lava balls does not vary with the position of the thin section, but that of the brown inclusions does.

In the section parallel to the tabular jointing the inclusions are, as a rule, round, oval, or trilobate, grouped either haphazardly or radially (Figure 6); in sections normal to this section, the rounded bodies are often arranged like a rosary, continuing as solid bands and locally protruding from the balls (Figure 5). This indicates that some of the drop-like bodies merge to form layers parallel to the tabular jointing planes.

Measurements of two different components of these "balls" in three mutually perpendicular

FIGURE 5. Various forms of brown bodies in felsitic groundmass and the isotropic cement glass

Magnification 15X; Nicols crossed.

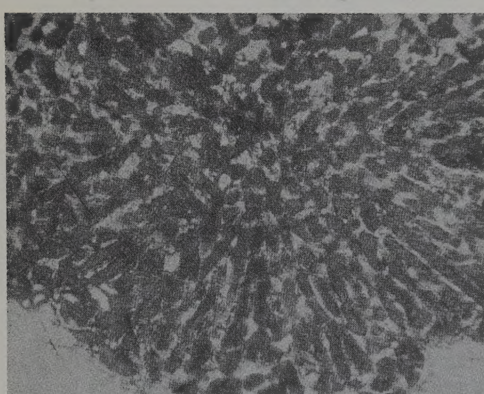


FIGURE 6. Various forms and distribution of brown bodies within a sphere

Section coinciding with tabular parting. Magnification, 7X; without analyzer.

sections give the following volumes, in % (Table 1):

It appears, then, that the number of brown bodies is greater in planes parallel to the tabular jointing than in planes normal to it; th seems to be due to the above-mentioned asso ation of the former with the individual brown glass "layers".

Also seen under the microscope are the following interesting features of the "balls":

a) together with the round sections are tho having one or more rectilinear boundary lines which coincide with the joints (Figure 7);

b) within individual "balls" the colorless glass cement is preserved in isolated rounded segments bounded by perlite jointing, with the yellow felsitic mass of the balls penetrating t colorless isotropic glass (Figure 8);

c) segments of felsitic substance, yellow in translucent light, locally occur outside the balls; they fringe individual joints or occur at their intersections (Figure 9);

d) while the boundary between the brown

Table 1

Section No.	Brown tear-shaped bodies	Felsitic groundmass	Remarks
B-12	50.2	49.8	Section parallel to tabular jointing
B-13	40.4	59.6	Section normal to tabular jointing
B-14	38.5	61.5	Section normal to tabular jointing

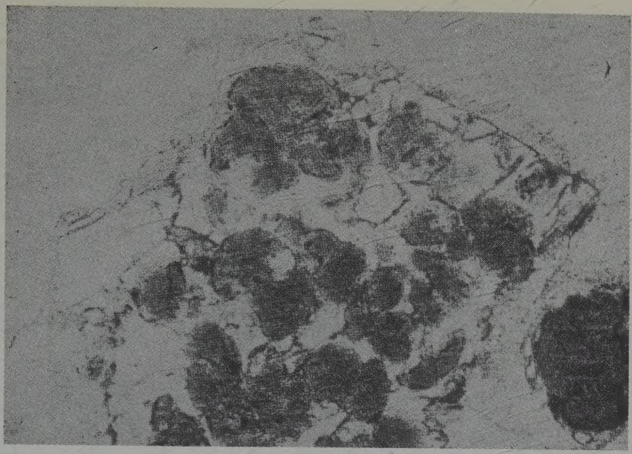


FIGURE 7. Rectilinear boundaries of some felsitic areas
Magnification 16X; without analyzer.

FIGURE 8. Secondary alterations along perlite jointing
Magnification 16X; without analyzer.

.es and the colorless cement glass is always rp, that between the glass and yellow felsite ften indistinct (Figure 5).

Ve believe that this definitely suggests that yellow felsite is a product of secondary ration of the colorless glass during the -magmatic crystallization stage. In other ds, we have here a miniature copy of what . Favorskaya [9] has observed in the nnyy Point obsidian on the Japanese Sea st. In both instances, secondary alterations iis type are mainly associated with the ıdaries between heterogeneous bodies — the tear-drop-like brown inclusions, in this instance. This explains the predominantly spherical shape of the altered portions.

We turn now to certain features suggestive of the manner of origin of these brown bodies. Of these features, the following are noteworthy:

a) banded bodies bend about the andesine incrustations:

b) the thin fibers, formed during crystallization, exhibit a linear arrangement subparallel to the axes of the above-mentioned

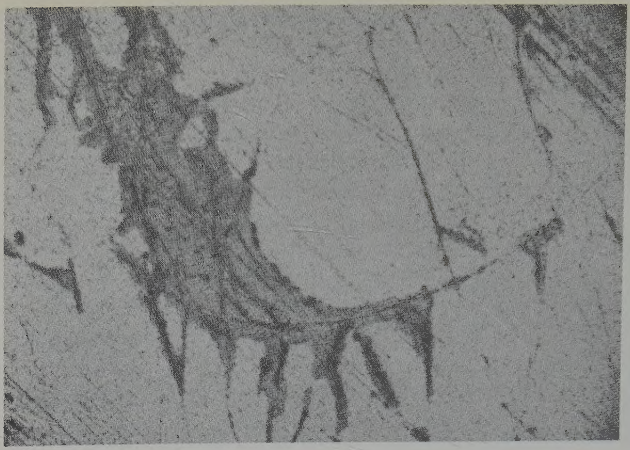


FIGURE 9. Secondary alterations in glass associated with joint intersections

Magnification 8X; without analyzer.

bands. This orientation is generally preserved even when the axial part of a band is interrupted by a joint or is associated with a crystal aggregate. Only rarely are the fibers oriented normal to the axis of a joint;

c) the arrangement of the above-mentioned crystallites is of interest. They are represented by the finest of acicules of magnetite and apatite arranged roughly parallel in the colorless glass. They disappear at the boundary with altered felsitic segments, and the felsite is almost completely free of inclusions. In contrast, at the boundary between the colorless glass and the brown bodies the crystallite chains either continue in the latter in a straig line, ignoring the boundary, or bend around th perlite joints and enter a brown body at an angle to their former direction (Figure 10). Some inclusions of the banded type are enrich in crystallite along the axes.

All of this suggests that the brown bodies were formed as the result of the liquation of a originally homogeneous melt.

One of the thin sections cut across a ball was dyed with cobalt nitrate, to demonstrate that the tear-like brown bodies are low in potassium while the secondarily altered

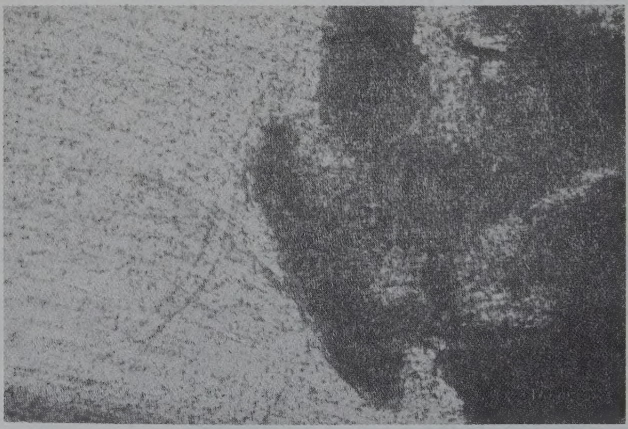


FIGURE 10. Crystallites at the boundary between colorless glass and brown bodies

Magnification 56X; without analyzer.

Elements	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	H_2O^-	H_2O^+	CO_2	F	B	P_2O_5	Totals
Volcanic glass	70,40	0,05	12,32	0,97	0,96	0,06	0,19	1,14	4,44	2,46	3,20	4,51	—	—	—	—	100,21
"Balls"	70,48	0,02	13,13	1,20	0,63	0,03	0,19	1,32	3,34	5,28	1,20	1,60	1,70	0,08	—	0,04	100,21

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felsitic segments are largely orthoclase feldspar.

Chemical analyses show that the composition of the balls differs from that of their glass (Table 2), the most substantial difference being in the alkali ratio and the constitutional water content. Inasmuch as it was impossible to isolate the brown bodies, they were analyzed together with the surrounding felsitic mass consisting mostly of orthoclase. This explains the high K_2O content in the balls. At the same time, there is more water in the colorless glass cement than in the balls where some of it was possibly lost during crystallization.

This detailed description of the Brusilovsk lava balls in the Imanka basin suggests that they originated as the result of two consecutive phenomena: liquation of a silicate melt and a subsequent action of late magmatic solutions on the solidified cement-glass.

The sequence of events was as follows: the melt was differentiated into two immiscible fluids; the tear-drop-like inclusions were probably the first to solidify; as a result, conditions favorable for the development of perlite jointings prevailed during the crystallization. Subsequently, crystallites formed in both the cement-glass and the solidified droplets; their arrangement was parallel to the prevailing trend of the joints — the horizontal — and the perlite, to a certain extent. Later, during the post-magmatic stage, potassium-rich solutions filled the joints and altered the glass to felsite and dissolved the crystals.

There is, however, another important aspect of this phenomenon. When the variolites are crystallized to some extent, it is often difficult to tell whether we are dealing with spherulites which were crystallized from a uniform primary melt or with local secondary alteration of a homogeneous primary glass. Such doubts were entertained by most scientists studying these ball-like formations in lavas. Thus, according to D. S. Belyankin [1], liquation can be postulated only for poorly crystallized balls which do not exhibit a radial structure.

There also is a difference of opinion between T. L. Tanton [11] and J. W. Greig [10], on the origin of these spherical bodies in the Agate Point quartz porphyries on the north shore of Lake Superior, Canada. Tenton holds that these lavas were formed during liquation; Greig, investigating these phenomena in quartz porphyries, believes them to be the product of common spherulitic devitrification.

Thus a demonstration of the presence of liquation calls for a careful selection of evidence in each particular instance.

Returning to the consideration of the Imanka

basin lava balls, the following facts support the hypothesis of their liquation origin:

a) the brown droplets have smooth and sharp boundaries with the enclosing glass; their component crystals never extend beyond the rounded outlines;

b) these droplets undoubtedly were liquid to start with, because they locally merge to form thin layers parallel to the tabular jointing, and are band-like in section;

c) that the brown bodies were not formed during the growth of spherulites out of a melt, and are not products of secondary alteration of the glass, as are the above-mentioned felsitic segments, is further attested by the behavior of the above-described crystallites at the boundary between these bodies and the glass; furthermore, the arrangement of the droplets' crystallization products precludes their growth out of a definite center; only in rare instances are they oriented normal to the joints — possible conduits for post-magmatic solutions.

On the contrary, the crystal orientation is, as a rule, almost linear and sub-parallel, the same as for individual groups in the adjacent bodies. This orientation corroborates what we have stated before to the effect that the droplets formed during liquation solidified before the surrounding melt; otherwise, the crystals would have grown from the periphery toward the centers of these bodies.

The cause of the stratification in the silicate melt, in this instance, can only be guessed at. Most students of the liquation of lavas see its causes in the abundance of volatiles, in particular, of water. This has been corroborated by D. P. Grigor'yev's experiments [4].

As mentioned before, it was impossible to analyze separately the drop-like bodies — the varioles, strictly speaking — so that we can only compare the composition of the cement-glass and that of the lava balls which contain both the varioles and the enclosing felsitic mass.

An attempt to calculate the variole composition from the quantitative analysis above, and assuming that the felsite consists of orthoclase feldspar, has shown that the K_2O content in the "balls" should have been much higher than established by the analysis. Consequently, albite and quartz are also probably present in the felsite along with orthoclase; their quantitative ratios cannot be determined. Withal, the chemical data thus obtained suggest that the presence of water in the glass is the latter's principal difference from the varioles. This again suggests, as in instances described in literature, that the high water content was the cause of liquation.

In conclusion, it should be noted that the liquation phenomena appear to be considerabl[y] more common in the crystallization of acid lavas than is assumed to be the case; howeve[r] their presence is not always readily demonstrable because of the devitrification and me[ta]somatism phenomena. It is not impossible th[at] these felsites containing the vermicular inclusions from the Brusilovsk formation of the Imanka River area, are also a product of liquation and of subsequent felsitic crystallization of the glass.

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Received, 3 July 1960

PHASES AND FACIES OF ALKALIC INTRUSIVES IN THE KHODZHAACHKAN RIVER BASIN (ALAY RANGE) IN CONNECTION WITH THEIR ORIGIN[1]

by

L. L. Perchuk, B. I. Omel'yanenko, and N. F. Shinkarev

The Lower Permian intrusive complex is the most widespread in the Turkestan-Alay. As of now, this complex is subdivided into the following three non-contemporaneous groups (from older to younger):

1. Granodiorites and quartz-diorites in stocks covering a few to tens of square kilometers;

2. High alkalinity porphyritic granites occurring in comparatively large (over 100 sq km) intrusive massifs;

3. Alkalic rocks which form intrusive massifs not exceeding a few square kilometers in area. There are more than a score such intrusives in the Turkestan-Alay.

Although these alkalic massifs long ago attracted the attention of scientists, they are comparatively little known as yet, largely because of their inaccessibility.

It was believed, until recently, that the formation of this Lower Permian intrusive complex culminated in the intrusion of nepheline syenites. Our own field work during the last four years has demonstrated a multi-phase origin of some of the alkalic massifs and has disproved the nepheline syenite stage as terminal [9, 14]. It also made it possible to review some aspects of their origin.

STRUCTURAL GEOLOGY AND AGE OF THE ALKALIC INTRUSIVES

The intrusives of the Khodzhaachkan River basin are located at the junction of two structural and facies zones: the Surmetash-Khodzhaachkan synclinorium and the Turkestan-Zeravshan anticlinorium (after G. S. Porshnyakov and D. P. Rezvyy). Only small portions of these zones fall on the geologic map (Figure 1); they are separated by sublatitudinal regional faults. South of one such fault, within the Turkestan-Zeravshan anticlinorium, there are a number of second-ord folded structures, consisting of Silurian calcareous-arenaceous-argillaceous deposits, w three transitional formations: from the arenaceous-argillaceous Llandoverian — through the Wenlockian calcareous shales — to the Ludlovian shales. Farther south there is a Lower- to Middle Devonian narrow, sublatitudinal graben. North of the regional fau the extensive Upper Carboniferous shales are transgressively overlain by Upper Carbonifer to Lower Permian conglomerates.

The obvious association of the intrusives w this regional fault gives reason to postulate th the fault line served as a magma conduit, with type of the intrusion being determined by elements of the enclosing folded structures. Th the Kul'p intrusive occurs in the center of an anticlinal fold and is stock-like in shape. Th Khodzhaachkan massif which fills the fault where it bends sharply, is asymmetrically teardrop-like. The Dzhilisuy massif, in the center of a synclinal fold, is pear-shaped.

The age of these alkalic intrusions is supposed to be, quite tentatively, Lower Permian Their absolute age, as determined by the K-A method on mica from the Kul'p massif pegmati in the laboratory of the All-Union Geological I stitute (VSEGEI), is 237 million years. As de termined by the lead method on three hatchett lite samples from the Khodzhaachkan massif, by I. D. Bespalova, at the Institute of the Geology of Ore Deposits, Petrography, Mineralog and Geochemistry (IGEM), it is 210, 217, and 220 million years, respectively. These figur together with the available geological data, ar enough to consider that these alkalic intrusive are Lower Permian in age.

PRINCIPAL STRUCTURAL FEATURES OF TH KHODZHAACHKAN BASIN ALKALIC INTRUSIVE MASSIFS

These rocks were formed in three consecu

[1]Fazy i fatsii shchelochnykh intruzivov basseyna r. Khodzhaachkan (Alayskiy Khrebet) v svyazi s voprosami ikh genezisa, (pp. 13-23).

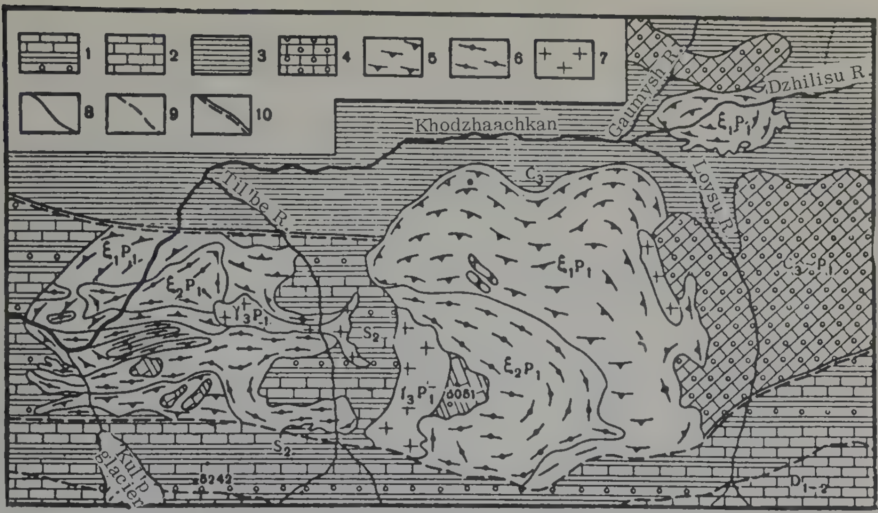

FIGURE 1. Geologic structure map of the Khodzhaachkan River basin
Scale, 1:130,000

1 - Silurian calcareous-arenaceous-argillaceous deposits (S_2); 2 - Devonian limestones (D_{1-2}); 3 - Upper Carboniferous shales (C_3); 4 - calcareous conglomerates, undifferentiated Upper Carboniferous-Lower Permian (C_3 - P_1); 5 - nepheline syenites (phase one); 6 - syenites and quartz syenites (phase two); 7 - leucocratic granites (phase three); 8 - stratigraphic and intrusive contacts; 9 - faults; 10 - regional fault.

usive phases, as follows (from older to nger): 1) biotite nepheline syenites and r facies; 2) biotite quartz syenites and their es; and 3) biotite granites. These rock s are readily distinguished in the field by r sharp contacts, elements of internal icture (trachytic aspect, linear structure,), and petrographic composition.

Table 1 reflects the geology and composi- s of the several phases and facies in the dzhaachkan River basin massifs. It shows the rocks of all of the intrusive facies are sent in the Khodzhaachkan and Kul'p mas- , but the Dzhilisu massif is made up exclu- ely of first-phase nepheline syenites. For of space we omit the detailed description ll rock phases, save for facts having a ect bearing on their origin. The quantitative- eral and chemical compositions of the ious facies are given in Tables 2 and 3. The ivatives of these facies within the intrusive ssifs are shown on the geologic map (Figure

These data, in conjunction with those pub- ed previously [1, 2, 8-11, 13, 14], make ossible to consider a number of questions ated to the origin of the phases and facies lkalic intrusions.

ORMATION CONDITIONS OF THE PHASE ONE FACIES

As shown in Table 1, rocks of the first intrusive facies are represented chiefly by nepheline syenites. They are represented by six facies, as follows.

A. Biotitic nepheline syenites, most common in the Turkestan-Alay, are developed in all three massifs. Being emplaced in the centers of these massifs, they do not exhibit as a rule any essential features of the assimilation of the enclosing rocks by the alkalic magma. There are reasons for assuming that the composition of these syenites is, on the whole, that of the intruded magma. Present instead of nepheline in some varieties are cancrinite, sodalite, and liebenerite associated with a post-magmatic alteration of the nepheline.

B. Aegirine-augite- and amphibole-nepheline syenites occur in all three massifs but are most extensive in the Khodzhaachkan massif where they cover over half of the area. A typical feature of their mineral composition is the considerable variation in the content of the light minerals. The amount of aegirine-augite and amphibole in the melanocratic varieties often exceeds 40% and amounts to 6-8% in the leucocratic varieties. Aegirine-augite predominates in some instances, and amphibole in others. At times both minerals are present in about the same amounts, with aegirine-augite often altered to amphibole.

In a number of instances, pyroxene-amphibole nepheline syenites are present at the contact of nepheline syenites and limestones; in

Table 1

Differentiation of alkalic intrusives in the Khodzhaachkan basin

Intrusive phases and their composition	Morphological types of intrusive bodies	Intrusive facies and their composition	Geologic occurrence of facies	Distribution of facies by the massifs
Third intrusive phase (leucocratic granites)	Dike-like bodies up to 700 m thick and up to 2.5 km long. Thin veins and dikes	Fine-grained biotite and tourmaline granites	In rocks of facies one and two and in the enclosing rocks	Kul'p and Khodzhaa-chkan
Second intrusive phase (trachytoid alkalic syenites and quartz syenites)	Isolated large stock-like and dike-like bodies. Thin veins and dikes	A. Biotite quartz-syenites	Comprise about 50% of intrusive bodies, mostly dikes and veins. Occur largely in central parts of intrusions	Kul'p
		B. Aegirine-augite- and quartz aegirine-augite syenites	In contaminated and contact sections of second-phase intrusions	Kul'p and Khodzhaa-chkan
First intrusive phase (coarse-grained nepheline and alkalic syenites)	Isolated, large stock- and dike-like bodies in the Kul'p massif comprise up to 75% of the Khodzhaachkan massif and all of the Dzhilisu massif	A. Biotite nepheline syenites	Comprise about 30–40% of phase-one bodies, mostly in their central parts	Kul'p, Khodzhaachkan and Dzhilisuy
		B. Aegirine-augite- and amphibole nepheline syenites	Comprise large areas of about 50–60% of phase-one rocks; often associated with peripheral parts of intrusives but also occur in their central parts;	Same
		C. Biotite-amphibole nepheline syenites	Form bands in the biotite nepheline syenites; also occur at contacts with the enclosing Silurian shales and their xenoliths	Kul'p
		D. Amphibole-biotite alkalic syenites	Form bands in the biotite nepheline syenites; also occur at the contact of first-phase intrusions and Silurian shales and their xenoliths	Kul'p
		E. Alkalic hybrid rocks	At contacts with Silurian and Permo-Carboniferous rocks and skarns	Khodzhaachkan and Kul'p
		F. Nepheline syenite-gneisses	Located largely near the top where they form motley bands in shales	Khodzhaachkan

Table 2

Qualitative-mineralogical composition of akalie rocks in the Khodyhaachkan River basin (% by volume)

Phases / Facies / Minerals	Nepheline syenites (phase one)							
	biotite		aegirine-augite- and amphibole		biotite-amphibole- and nepheline-bearing		Amphibole-biotite, nepheline-free	
	Variation range	Average of [11] samples	Variation range	Average of [11] samples	Variation range	Average of [11] samples	Variation range	Average of [11] samples
Microcline	14.2—58.4	44.0	6.2—68.1	45.6	59.5—61.4	60.4	57.8—65.1	62.6
Albite	5.9—53.3	24.0	6.6—30.2	11.8	10.8—14.9	14.9	5.8—28.6	15.7
Quartz	—	—	—	—	—	—	—	—
Nepheline	2.9—47.3	18.9	5.5—23.5	17.8	1.5—5.4	3.8	—	—
Biotite	0.6—11.5	4.5	0.0—2.1	0.5	2.2—10.2	7.4	0.0—8.3	3.4
Amphibole	—	—	0.0—22.1	6.5	4.3—12.3	8.5	2.1—26.3	15.2
Aegirine-Augite	—	—	0.0—18.2	6.3	0.0—1.2	0.8	—	—
Calcite	0.0—5.8	0.8	0.0—3.5	1.5	—	—	0.0—2.8	1.0
Granite	0.0—5.0	1.4	0.0—4.8	0.8	—	—	—	—
Sphene	0.0—4.3	0.7	0.2—1.5	0.9	2.4—4.0	3.1	0.2—2.6	2.1
Wollastonite	—	—	—	—	—	—	—	—
Cancrinite	0.0—9.8	2.7	0.0—4.5	1.2	0.0—1.2	0.3	—	—
Spreustein	0.0—19.4	2.0	0.0—37.6	7.3	0.0—2.4	0.8	—	—
Sodalite	0.0—4.2	1.0	—	—	—	—	—	—

Table 2 (continued)

Phases / Facies / Minerals	Quartz syenites (phase two)										Remarks
	biotite		aegirine-augite- and amphibole		biotite-amphibole- and nepheline-bearing		amphibole-biotite, nepheline-free		biotite		
	Variation range	Average [11] samp.	Variation range	Average [11] samp.	Variation range	Average [11] samp.	Variation range	Average [11] samp.	Variation range	Average [11] samp.	
Microcline	60.1—63.0	61.5	45.7—58.3	52.0	27.3—58.8	50.0	17.5—28.4	18.3	35.4—46.5	42.6	
Albite	8.3—14.8	11.5	15.4—29.2	22.3	7.3—38.5	17.5	—	—	17.0—30.0	24.3	Secondary in syenites
Quartz	20.9—21.0	20.6	0.0—8.2	4.1	0.0—11.3	7.6	—	—	24.0—31.1	28.3	—
Nepheline	—	—	—	—	—	—	8.0—9.9	3.4	—	—	—
Biotite	40.—5.6	4.8	4.8—11.0	7.2	—	—	—	—	3.1—6.9	4.8	
Amphibole	—	—	5.6—18.8	12.2	0.0—1.1	0.4	—	—	—	—	Largely hastingsite
Aegirine-Augite	—	—	—	—	18.7—28.5	20.8	45.0—48.9	48.0	—	—	—
Calcite	—	—	—	—	—	—	1.3—12.9	12.0	—	—	In assoc. magnetite
Granite	—	—	—	—	0.0—4.3	1.4	10.9—12.3	11.1	—	—	—
Sphene	0.2—2.1	1.6	1.5—7.0	2.2	1.3—3.2	2.0	0.0—1.5	1.0	—	—	In assoc. with zircon
Wollastonite	—	—	—	—	0.0—1.1	0.3	0.0—6.9	6.0	—	—	—
Cancrinite	—	—	—	—	—	—	—	—	—	—	—
Spreustein	—	—	—	—	—	—	—	—	—	—	—
Sodalite	—	—	—	—	—	—	—	—	—	—	—

Table 3

Chemical composition of rocks from the Khodzhaachkan intrusive complex

Oxides	Nepheline syenites (phase one)					Quartz syenites (phase two)			Granites (phase three)	
	biotite		aegirine-agite- and amphibole			biotite	pyroxene	hybrid alkalic rocks	biotite	
	1	2	3	4	5	6	7	8	9	10
SiO_2	53.76	56.25	58.65	54.42	54.00	69.75	61.49	44.99	72.97	72.92
TiO_2	0.38	0.36	0.27	0.14	0.70	0.38	0.38	1.95	0.32	Trace
Al_2O_3	22.64	23.39	17.64	22.20	20.45	14.75	15.10	4.72	13.72	14.00
Fe_2O_3	0.54	0.11	1.18	0.65	1.27	0.32	2.30	10.16	0.12	0.10
FeO	3.87	1.69	3.20	3.01	4.62	2.88	2.87	8.21	1.76	1.76
MnO	0.09	0.11	0.14	0.11	0.13	0.07	0.22	1.23	0.04	Trace
CaO	2.47	1.05	2.95	3.03	5.46	2.45	5.45	22.29	1.37	1.33
MgO	0.47	0.07	0.96	0.14	0.70	1.04	1.14	2.09	0.90	0.35
K_2O	7.24	4.92	5.90	8.05	5.46	3.52	5.53	1.64	5.34	4.81
Na_2O	6.67	10.02	6.35	6.89	6.29	3.81	5.37	2.05	2.81	3.45
P_2O_5	0.05	—	0.01	—	0.13	0.14	0.09	0.33	—	—
CO_2	—	0.42	—	0.30	0.50	—	—	—	—	—
F	—	—	0.37	—	—	0.29	0.18	0.26	—	—
H_2O^+	—	0.80	—	0.22	—	—	—	—	0.06	—
Losses in heating	2.05	0.60	2.89	0.11	0.69	0.57	0.21	0.37	0.80	0.97
Total	100.23	99.79	100.51	99.27	100.40	99.97	100.33	99.99	100.21	99 69

Analyses 1, 2, 4, and 5 performed in the chemical laboratory of the IGEM of the Academy of Sciences, U.S.S.R.; analyses 3, 6, 7-10 — in the laboratory of the Institute of Silicate Chemistry, Academy of Sciences, U.S.S.R.

such instances the assimilation of limestones by an alkaline magma is obvious. On the other hand, pyroxene-amphibole nepheline syenites often comprise large sections in the biotitic nepheline syenites. In such instances, leucocratic and mesocratic varieties predominate and the rock composition is relatively constant over considerable areas. A comparison of the chemical composition of the biotitic and aegirine-augite nepheline syenites (Table 3) shows the latter are characterized by a higher calcium oxide content. This suggests that the alkaline magma, reacting with deep-seated carbonate rocks (along its intrusion path), became enriched in calcium oxide to become pyroxene and amphibole instead of biotite. It is of interest that the aegirine-augite nepheline syenites are comparable in chemical and mineral composition to the nepheline syenites of the Botogol Knob with its extensively developed reactions with the limestones.

C and D. Biotite-amphibole syenites with or without nepheline. These rocks have been observed only in the Kul'p massif where they form thin bands in the biotite nepheline syenite or occur at endocontacts of the latter and the Silurian shales. This facies is often replaced by nepheline-free syenites where shale xenoliths are developed. Gradual transitions from biotitic nepheline syenites to nepheline-bearing and then to nepheline-free, are present almost everywhere. The decrease in nepheline content is accompanied by an increase in primary albite, amphibole, and partly in microcline. The "de-nephelinization" reaction apparently proceeds as follows,

$$\underset{\text{nepheline}}{NaAlSiO_4} + \underset{\text{quartz}}{2SiO_2} = \underset{\text{albite}}{NaAlSi_3O_8}$$

and is related to the assimilation of the Silurian arenaceous-argillaceous deposits by biotite nepheline syenites. These facies have not been observed at the nepheline syenite and limestone contacts.

Although these rocks account for a very small portion of the total area, they indicate the extreme diversity in the composition of the alkaline magma, depending on the composition of the contaminating rocks.

E. Hybrid alkalic rocks associated with the first-phase intrusions are quite diversified in

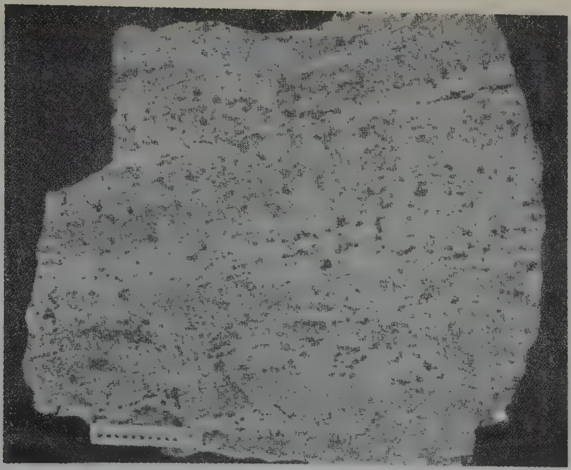

FIGURE 2. Coarsely-banded nepheline syenite-gneiss.

nposition and structure. They always occur areas where there are limestone xenoliths d at direct contacts between the nepheline enites and the enclosing carbonate rocks. oskarns are often present at the boundary ween the hybrid alkalic rocks and the limenes; they consist of aegirine-augite, wollanite, garnet, and calcite. The principal nerals of the hybrid rocks are aegirinegite, garnet, microcline, calcite, nepheline, d sphene. The quantitative ratios of these nerals are quite inconsistent (Table 2), with girine-augite and microcline predominating a rule. The high aegirine-augite content and presence of some calcite, garnet, and casional wollasonite are the specific features these rocks. The presence of transitions om the hybrid igneous alkalic rocks to the nestones by way of metasomatic calcitecrocline-pyroxene rocks, suggests that ey originated in a diffusion magmatic replacement (assimilation) of limestones by the pheline-syenite magma.

F. Nepheline syenite-gneisses are distribud in the upper part of the Khodzhaachkan assif. They are marked by a distinct banding igures 2 and 3) caused by alternating darker ne-grained and lighter medium-grained rocks. ne bands vary in thickness from a few millieters to a few tens of centimeters. When the ernating bands are thin, their boundaries are gue. These rocks are mostly mesocratic girine-augite nepheline syenites, with the rker bands containing somewhat more girine-augite. The nepheline syeniteeiss sections contain numerous shale noliths, elongated with the banding. As a le, these xenoliths are recrystallized, with ne-grained aegirine-augite nepheline syenite the peripheral parts of the larger xenoliths similar to the dark bands in the nepheline syenite-gneisses. Closer to the xenolith center there are rocks consisting of quartz, microcline, aegirine-augite, and arfvedsonite; farther on, these change to quartz-diopside-plagioclase hornfels and then to shales. In steep canyon wall exposures, it can be seen that the xenoliths in the nepheline syenite-gneisses become more numerous higher in the section, while the syenite-gneisses gradually change to hornfels and shales intensely injected with numerous apophyses of nepheline syenites.

All this suggests that the nepheline syenite-gneisses are the result of the infiltrating magmatic replacement of shales by the nepheline-syenite magma.

THE FORMATION CONDITIONS OF SECOND-PHASE FACIES VARIETIES

The quartz syenites of the second phase are represented by leucocratic rocks with a comparatively high quartz content. Their colored mineral is biotite. Sizable bodies are formed by three biotite quartz syenites in the Khodzhaachkan and Kul'p massifs. The rocks are characterized by a comparatively leucocratic aspect and a fairly consistent composition. In many instances there are sharp intrusive contacts with first phase-one rocks, where the nepheline syenites are cut and metamorphosed by the quartz syenites. All this suggests that biotite quartz syenites correspond, on the whole, to the second-phase magma composition. The transition from this to other facies is accompanied by a reduction in the quartz content (until its complete disappearance) and by an increase in the amphibole and pyroxene content. A study of their chemical analyses (Table 3) shows an increase in total alkalinity and a reduction in acidity from the biotitic quartz syenites to the aegirine-augite syenite facies.

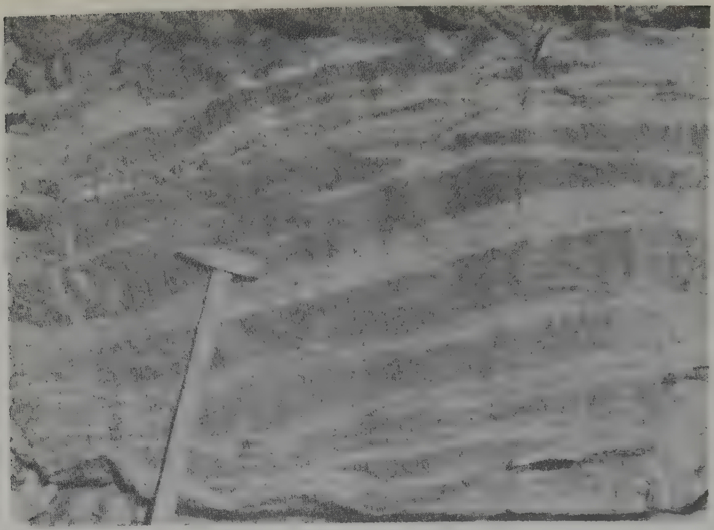

FIGURE 3. Thin-banded nepheline syenite-gneiss.

Aegirine-augite and quartz-bearing aegirine-augite syenites are especially widespread in the Khodzaachkan massif where they account for about 50% of the total area of the second-phase area studied. The numerous xenoliths of the enclosing rocks and the presence of large roof pendants of the uppermost Silurian limestones and shales show that we are dealing here with the upper parts of the intrusion. Contamination phenomena are also conspicuous in the aegirine-augite syenites in the Kul'p massif. Characteristically, hybrid alkalic rocks containing a considerably higher aegirine-augite content occur toward the limestone contact, along with calcite, garnet, nepheline, and sometimes wollastonite. Thus, the association of aegirine-syenites with the assimilation of carbonate rocks by the quartz-syenite magma is fairly obvious. The biotite-amphibole quartz syenites whose mineral composition is given in Table 2, were formed intermediately between biotite quartz syenites and the aegirine-augite syenites.

ROCKS OF THE THIRD INTRUSIVE PHASE

These rocks are represented by fine-grained leucocratic biotite granites. They form dike-like bodies up to 700 m thick and up to 2.5 km long. Usually they are thin. Tourmaline is common and abundant in these granites where it forms irregular aggregates up to 4-5 cm. Fine-grained tourmaline often fills the joint planes in the granites and is also developed in the enclosing rocks. All this indicates a post-magmatic autometasomatic origin of the tourmaline and explains the formation of the tourmaline granites.

* * *

This analysis of the formation conditions of the varieties of facies demonstrates that the composition of alkalic intrusions is closely associated with the differences in the primary composition of magmas in the various phases and with processes of assimilation of the enclosing rocks by the magma. We have considered only the most typical facies. Of course, many intermediate rocks are present in the massifs.

It should be added that the presence of such facies as liebenerite-, cancrinite-, and sodalite syenites and albitized nepheline syentites is associated with post-magmatic alterations of the first-phase nepheline syenites.

POSSIBLE CAUSES OF THE CHANGES IN COMPOSITION OF MAGMAS OF VARIOUS PHASES

The sequence of intrusive phenomena observed in the Turkestan-Alay Lower Permian intrusive complex — granodiorites and quartz diorites — porphyritic granites — alkalic rocks — suggests that the standard course of differentiation from more basic to more acid rocks took a turn, at a certain stage, toward the formation of alkalic magmas. The alkalic intrusions were formed in three phases: nepheline syenites — quartz syenites — leucocratic granites — with each successive phase more acid than the preceding. It is of interest to consider the reason for the observed change in content in the magmas of the successive phases.

is quite obvious that the problem of these anges is most closely related to that of the igin of alkalic rocks in general. There are any opinions on this subject. We shall conler it in the light of our earlier hypothesis].

Our explanation of the formation conditions r the alkalic magmas is based on the follows premises stated by D. S. Korzhinskiy in a mber of his works [3-7].

1. Granitization, in the broad sense, is imssible without the transition of the primarilydimentary rocks through a molten state. iis transition is preceded by a series of exange metasomatic reactions between the solid ck and the primary magmatic solution, where the rock first changes its composition, then comes a melt (magmatic replacement).

2. When a strong base and weak acid salt ıch as carbonates) are dissolved, the acidity the solution is reduced while the base tivity factor — especially for such strong ses as oxides of potassium and sodium — is creased. This premise has been theoretically bstantiated by D. S. Korzhinskiy [6].

According to the hypothesis, alkaline magma n be formed in the peripheral part of a anitic magmatic center in contact with carnate rocks. As the result of a magmatic reacement of these rocks in the primary magatic solution, there is a sharp increase in the ilcium and magnesium concentration. This, turn, leads to a growth in the activity factor, r all bases, and to an increase in the chemiil potentials for the solution alkalis. As a reilt of diffusion, potassium and sodium pass om the primary magmatic solution to the agma, while the reverse is true for silica. ius originates a higher alkalinity magma. asmuch as the diffusion of components from ie system to the other is determined by the emical potential difference, potassium and odium are displaced in a direction opposite to e solution movement. Because of this, those ortions of the granitic magma close to the agmatic replacement front also will be charterized by a higher alkalinity. In the magiatic chamber, the alkalinity will gradually ecrease, to the granite magma.

It is only natural that the most alkaline magıa is found in the peripheral part of the iamber. On these premises, the reduced lkalinity of the following phases is readily xplained. Indeed, nepheline syenites correond to the magma composition in the outer arts of the magmatic basin; quartz-syenites to a deeper part; with leucocratic granites orresponding to still greater depths. We beeve that the intrusion of the third-phase magıa took place during a comparatively quiesencent stage of tectono-igneous activity, with the magma issuing from the deeper portions of the chamber whose outer edges may already have been crystallized by that time.

This hypothesis easily explains the comparatively small size of the nepheline syenite massifs: with a greater volume of granite magma, the alkaline magma of the outer shell would have been completely dissolved in it.

Thus the magma composition of the various phases was determined by the structural pattern of the magma chamber, while the facies composition was determined largely by the processes of assimilation of the enclosing rocks by the alkaline magma — either in situ or along the intrusion path.[2]

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[2]This hypothesis is better understood if supported by geologic data; some of these data can be found in other articles by this author (see bibiliography). Editorial Office.

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Received, 15 February 1961

DOES FUEL GAS MIGRATE FROM THE DEEP STRATA IN THE KHIBINY MOUNTAINS?[1]

by

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ıe alkalic intrusive rocks of the Khibiny
if contain significant amounts of fuel gases
ich the saturated hydrocarbons (C_1 to C_4)
ominate. The nature of the gas show and
as saturation of the alkalic rocks have been
ribed adequately in the literature [2, 3,
].

ecently some scientists have expressed the
y that the presence of fuel gas in the
ny massif is associated with the migration
s along faults in the earth's crust which
d from the magma chamber to the earth's
ce. Thus, A.I. Kravtsov, in 1958 ex-
sed the theory that the fuel gases of the
iny Mountains were formed not only in the
ess of formation of this intrusion, but also
r as a migration from the depths to the
ce.

. 1956-60 we conducted a large number of
stigations on the deep migration of fuel gas
n the Khibiny massif. We studied the
ping surface gases, the presence of minute
tities of gas in the subsoil, the gas content
e joints and faults of intrusive rocks, the
s contained in the various alkalic rocks,
he inclusions in crystals. We also did
gas logging studies of some drill cores.

he presence of micro-gas shows in the sub-
was determined by boring holes 1.0 to 1.5
ep, sealing them hermetically by plugs,
taking samples of the subsoil air with a
r pump. The studies of free gases in the
s and fault zones of the intrusive rocks
conducted in apatite mine workings by
t holes 1.5 to 2.0 m deep, sealing them
netically, and taking samples of the gas by
ns of a water pump.

he gas samples were analyzed by modern
metric and chromatographic instruments,
ectrometer, and the widely used All-Union Technical Institute instrument. These instruments had a sensitivity of up to 0.0002 for hydrocarbon gases. 0.002 for hydrogen and carbon monoxide, and 0.1% for nitrogen and carbonic acid.

The few occurrences of escaping surface gas in the Khibiny massif (for example, the "Bolotnyi klyuch" site on the southern spur of Mt. Kukisvumchorr, etc.) proved to be air jets, basically of nitrogen and oxygen. No hydrogen and methane are present. Nowhere in the Khibiny region did we discover a surface outlet of fuel gases.

The microscopic subsurface gas shows were studied on a wide strip extending southward from the central parts of the massif to its contact with the Imandra-Varzuga sedimentary-effusives formation; that is, the gas content of all rock complexes of the intrusion was studied.

The fault zones overlain by thin morainic deposits (the Saamskaya, Gakman, and other river valleys) and the areas beyond them were studied.

A fault zone in a thalweg of the Saamskaya River valley was revealed by a drill core (L. B. Antonov, 1952); most investigators assume its connection with the faults of the other valleys studied.

The micro-study of subsoil gas showed that there is a slight migration of gaseous hydrocarbons (hydrogen is encountered only at a few points) from the gas-saturated intrusives into the atmosphere. This migration is not related to the fault zones. As a rule, there is a relatively low hydrocarbon content in soil gas above the fault zones. A distinct relationship is noted between the gas saturation of the intrusive rocks (the gas contained in rock interstices and in mineral cavities) and the fuel-gas concentration in the subsoil (Table 1).

Hence, the extremely small show of fuel gas in the subsoil is due to its diffusion from the gas-saturated intrusives through the thin Quaternary mantle into the atmosphere. No migration of fuel gas through the faults was observed.

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est' li potok goryuchikh gazov s glubin v
nakh?, (pp. 24-29).

Table 1

Relationship between the gas content of intrusive rocks and of subsoil air

Rocks	Average content of gaseous hydrocarbons in rocks, cm^3/kg	Content of gaseous hydrocarbons in subsoil air ($\% \cdot 10^{-5}$)			
		Number of points sampled	Variation of content		Average content
			Min.	Max.	
Foyaites	20.50	96	0.0	224	66.9
Ristochorrites	10.06	81	0.0	238	48.4
Apatite-nepheline ore body	3.90	16	14	140	63.9
Iolite-urtites	30.00	105	0.0	2324	119.0
Khibinites	48.80	106	0.0	1 0976	200,0
Hornfels (region of contact with rocks of the Imandra-Varzuga formation)	7.97	40	28	364	71.4

The soil-gas composition was as follows: CO_2, 0.1 to 13.88; O_2, 14 to 21.17; N_2, 71.39; He, 0.001; Ar, 0.85; CH_4, 0.0001 to 0.10976; C_3H_8, 0.00009; C_4H_{10}; 0.00005; H_2, 0.0 to 0.3.[2]

the massif in accordance with the laws of effusion, as distinguished from gases which ar adsorbed by the rocks and are contained in th enclosed interstices of the rocks and the mineral cavities.

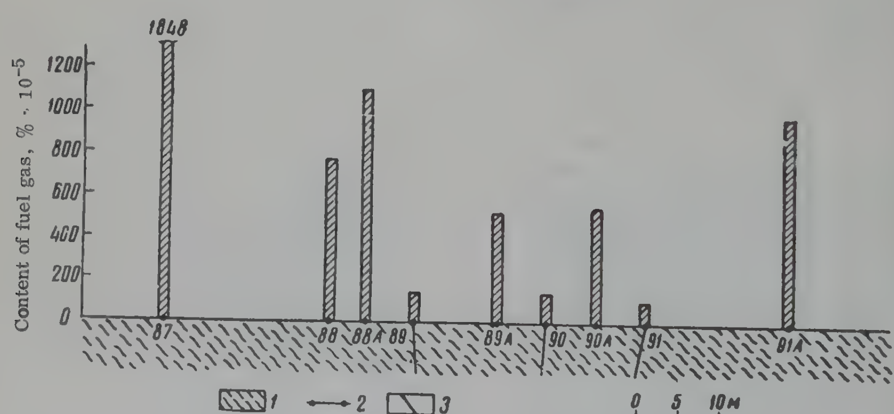

FIGURE 1. Content of free fuel gas in jointed and macroscopically non-jointed rocks at site 5, 404-m level of S.M. Kirov mine

1 - lenticular-banded apatite-nepheline rocks; 2 - boreholes from which samples of gas were taken; 3 - joints.

As indicated above, a study was made of the macro- and micro-fissures of the free gases in fault zones, in the large intersticies of the intrusives of the Khibiny apatite deposits.

We use the term "free gases" to designate those gases which migrate among the rocks of

The drill cores taken from the mine workin showed that considerably less fuel gas is contained in the joints cutting the alkalic intrusives than in the macroscopically dense rocks Thus, at site 5 (a mined working cutting the ore vein of the S. M. Kirov mine at the 404-meter level, the natural gas content in adjacent boreholes of the same rocks differs markedly depending on whether the boreholes are located in jointed or non-jointed rocks (Table 2, Figure 1).

[2] The presence of hydrogen was noted at only a few points.

Table 2

Fuel gas content of jointed and non-jointed rocks

No. of bore-hole	Fuel gas content $\% \cdot 10^{-5}$	Nature of rocks
87	1848	Macroscopically non-jointed
88	756	" "
88A	1120	" "
89	140	Gaping joint
89A	532	Non-jointed
90	140	Gaping joint, 3 cm wide
90A	560	Non-fissured
91	126	Joint filled with natrolite
91A	980	Non-jointed

he fault zones uncovered by mine work- are 10 to 30 m thick, and traceable for t distances. They are the youngest for- ons cutting the veined rocks. The pro- s of spreusteinization, zeolitization and nification are prevalent in these zones; is, the latter contain rocks which differ 1 the surrounding intrusive rocks in their adsorptive properties.

ll boreholes in the fault zones showed a fuel-gas content, not exceeding thousandths percent (Table 3).

comparison of fuel-gas content in a bore- produced in a fault zone uncovered by the spor tunnel and in a series of holes bored nse (macroscopically non-jointed) rocks interest:

No. of borehole	Fuel-gas content $\% \cdot 10^{-5}$	Rocks
15+37	6188	Pegmatite
15+44	224	Fault zone
15+55	2380	Urtite

The rocks filling the fault zones, while differing in their high adsorptive properties with respect to gas, contain almost no fuel gas. Thus, the fuel-gas content in saponite taken from the fault zone amounts to 0.0014 cm^3 per kg of rock, while in spreustein it does not exceed hundredths of a cubic centimeter. At the same time, the intrusive rocks surrounding the fault zones have a fuel-gas content of the order of tenths of a cubic centimeter per kg of rock.

Table 3

Sampling data on boreholes produced in the fracture zones

No. of bore-hole	Mined area	Fuel gas content $\% \cdot 10^{-5}$
	S.M. Kirov Mine	
48	Level + 404 m, field entry	84
53	" "	56
65	" "	74
	Yukspor Mine	
30	Level + 670 m, southwest entry	32
44	" "	36
Pk 17+10	Yukspor tunnel	28
Pk 15+44	" "	224

Thus, no fuel-gas migration is currently seen in the zones of the fractures which cut the alkalic rocks much later. Nor did such a migration occur earlier; otherwise, the fuel gas would have been detected in sorbed state in the "inclusions" in the rocks filling these zones. This serves as further confirmation of the opinion earlier expressed that the fuel gases of the Khibiny Mountains are syngenetic with alkalic rocks and were formed exclusively in the process of intrusion [2, 3].

Apart from the free gases, the fuel gases, chiefly gaseous hydrocarbons, are contained in the interstitial spaces in the alkalic rocks in an amount up to 240 cm^3 per kg of rock. If the thickness of the intrusive rocks comprising the massif is assumed to be 2.5 km, the gaseous hydrocarbons in the rock interstices alone amount to 200 to $250 \cdot 10^6 m^3$.

The various rocks differ in content of gaseous hydrocarbons (Table 1).

The study of the rock-forming minerals of the Khibiny massif has shown that the various minerals also differ markedly in fuel-gas content in the cavities (the inclusions) of the crystals. The average content of fuel gases in the cavities of the chief rock-forming minerals of the rocks exposed by the apatite mines is as follows:

Gas content in cm^3 per kg of rock

	Gaseous hydrocarbons	H	CO_2
Nepheline	18.00	0.29	1.04
Aegirite	6.30	0.23	0.0
Apatite	0.10	0.08	0.05

The high content of fuel gases enclosed in the pores of the rocks in the cross section exposed by the mine workings is characteristic of the rocks of the ijolite-urtite complex under the apatite-nepheline deposit and encountered among the apatite rocks in the form of xenoliths containing up to 82% nepheline, 6 to 8% aegirite, and 2% apatite. Rather large concentrations of fuel gases are noted in the rocks of the ore-poor zone of the deposit, represented by the lenticular-banded and sometimes blocky nepheline-apatite rocks containing 46 to 51% nepheline, 26.1 to 36% apatite, and 9 to 10% aegirite. In the mottled apatite-nepheline rocks, the rich zone of the deposit, which contain 64% apatite and 20 to 24% nepheline, minimum amounts of fuel gases not exceeding tenths of a cm^3 per kg of rock are observed.

The increased concentrations of free gases found in sampling the boreholes of the apatite mine workings were also associated with th ijolite-urtites. Somewhat smaller contents these gases were observed in the lenticular-banded nepheline-apatite rocks, and minimu contents close to the limit of sensitivity of th gas-analyzing apparatus used were found in rich zone of the deposit.

Thus it is possible to say that the content bound gas and free gas corresponds. The ga content in the free phase is related not to the fissure structure of the intrusive massif, bu to the gas saturation of the rocks. This is splendidly illustrated by the results of the bo hole sampling at site 42 at the 392 m level of the S. M. Kirov mine (Figure 2). Here the hi content of free gaseous hydrocarbons is associated with the lenticular-banded and blo structure rocks; almost no gaseous hydrocarbons are observed in the mottled apatite-nepheline rocks, but the content increases again markedly in the xenolite of the ijolite-urtite rocks in the ore-rich zone of the deposit.

In confirmation of the occurrence of the migration of fuel gas from the deep-lying str A. I. Kravtsov [1] refers to a small increase in fuel-gas content in the clay suspension in t lower part of the section of borehole 1-G, wh he does not attribute to geological conditions.

Borehole 1-G was drilled at the 92 m level of an apatite mine working to an absolute dept of 300 m. The upper part of the column, to a depth of 64 m, is composed basically of mott and block-shaped rocks rich in apatite. Fractured rocks and ijolite-urtites, in which nephe line predominates, were revealed beginning a 64 m to the base. The hydrocarbon-gas conte in the pores of the rocks along the column to depth of 64 m did not exceed 0.5 cm^3 per 1 kg while it was more than 4 times as great in the lower part of the column, in the ijolite-urtite The increase in gas content in the lower part the column is associated not with migration of gases from the deep strata, but with the presence of rocks of different composition having greater gas saturability.

The core sampling investigations we condu ted in uniform rocks (ijolites) at hole 168 (the Saamskaya River valley) down to 250 m showe no increase in fuel-gas concentration with dep

The increased content of free fuel gas in th apatite mines is associated with local section having an area of about 16,000 m^2 located in the ijolite-urtites and the lenticular-banded apatite-nepheline rocks. Zones of increased concentration of fuel gas differ locally with th reservoir properties of rocks of the same typ the permeability of which ranges from 0.001 0.44 millidarsy in rocks not macroscopically jointed. The more permeable rocks, enclose by practically impermeable rocks, present

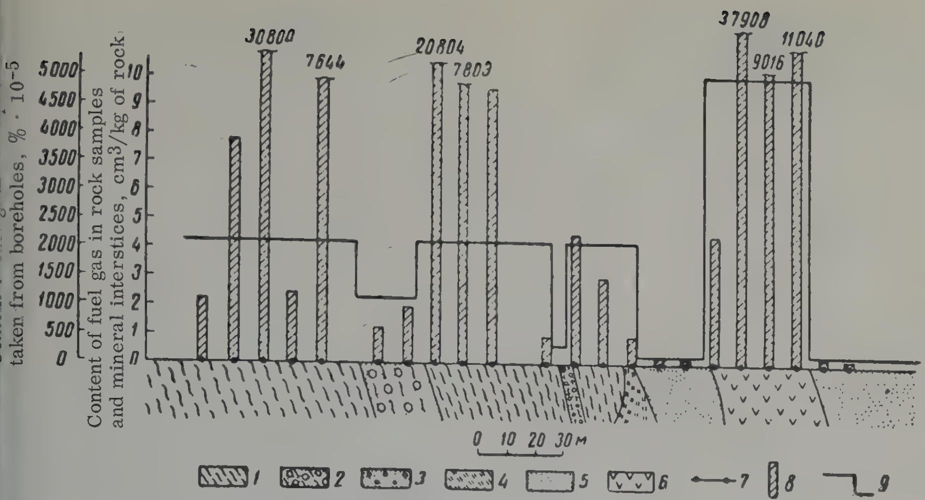

FIGURE 2. Content of free fuel gas and fuel gas in rock interstices and mineral cavities at site 42 of the 392-m level of the S.M. Kirov mine.

1 - lenticular-banded apatite-nepheline rocks; 2 - large-block apatite-nepheline rocks; 3 - small-block apatite-nepheline rocks; 4 - reticulate apatite-nepheline rocks; 5 - mottled apatite-nepheline rocks; 6 - urtites; 7 - boreholes from which gas samples were taken; 8 - content of fuel gas in rock interstices and mineral cavities.

rable conditions for the accumulation of fuel gas.

he varying reservoir properties of the lic rocks are an important factor in the ac- ulation of significant amounts of fuel gas in vidual locales. The content of the hydro- bon gases in these sectors is as high as the hydrogen content does not exceed %.

he free gas in the studied part of the mn within the 300-to-700 m level repre- s a mizture of the gas at depth with atmo- eric gas penetrating the massif along the erous joints in the rocks.

he composition of the free gas, according ie borehole sampling data is as follows, in cent based on volume: O_2, 15 to 20.9%; , 0 to 1.97%; CH_4, 0.00028 to 16.50%; 6, 0.00022 to 1.40%; C_3H_8, 0.00011 to 3%; C_4H_{10}, 0 to 0.158%; H_2, 0 to 0.70%; , 0 to 0.5%; N_2, 65.31 to 81.9%; Ar, 0.95 .98%; He, 0.0011 to 0.0013%; $\frac{Ar \cdot 100}{N_2 \cdot 1.18}$, 3 to 1.08%.

hese data show that there is no deep migra- of fuel gas from the interior of the Khibiny sif to the surface.

distinct relationship between the content ee fuel gas and the gas saturation of the s is evident. The increased concentrations of the fuel gas content of these rocks is associated with specific minerals and with the rocks in which these minerals predominate.

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Received, 27 June 1960

ON THE TECTONICS OF THE KUZNETSK ALATAU[1]

by

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he tectonics of Kuznetsk Alatau were
ed as far back as 1920-1930 by V. A.
chev, A. N. Churakov, Ya. S. Edel'steyn,
Bazhenov, A. M. Kuz'min, K. V. Radugin,
). V. Nikitin. In recent years it has been
ed by A. L. Dodin, V. A. Kuznetsov, T. M.
o, N. S. Zaytsev, V. A. Unskov, V. V.
nentovskiy, D. I. Musatov, A. P. Tarkov,
). Skobelev, and others.

ne concepts on the Kuznetsk Alatau tec-
s, as postulated in those works, are
contradictory. For instance, A. L.
n [3] believes that the principal down-
ing of the Caledonian geosyncline coin-
with the axial part of the range in its
eridional trend and that its structure is
ed by linear folds which spread out fan-
o the northwest and northeast.

A. Unskov [14] explains the intricate con-
ation and the great diversity in the strike
ctonic structures in the Kuznetsk Alatau
eing due to the rigid pre-Sinian basement
ected by crisscrossing faults into a com-
mosaic of block structures.

ccording to D. I. Musatov and A. P. Tarkov
, the structural features of the Kuznetsk
u are determined by the development of a
ow Riphean and Lower Cambrian geosyn-
l trough which crossed the range from
west to southeast, with a system of 15
easterly trending geosynclines and uplifts
e east of it.

. V. Khomentovskiy [13] associates the
etsk Alatau tectonic structures with the
ation of a deep-seated arcuate Riphean rift
lting in a wide zone of volcanic and meta-
phic rocks along its western slope.

. N. Krasil'nikov, A. A. Mossakovskiy [6],
others identify three principal structural elements on the basis of formation analysis: the North and South Minusinsk intrageosynclinal downwarps and the Batenevsk intrageosynclinal uplift, with each element having its own complex structure.

Just as controversial is the time of formation of the Kuznetsk Alatau. Some investigators believe it to be Caledonian [2, 3] formed on the Ordovician-Silurian boundary. Others believe that the folded structures of the Kuznetsk Alatau, Gornaya Shoria, the southwestern part of Eastern Sayan, and the northern part of Western Sayan, are late Cambrian, contemporaneous with the Cambro-Ordovician "Salairian" folding of South Siberia. This interpretation of the tectonic structures of the Kuznetsk Alatau is based on the work of V. A. Kuznetsov [7] and is reflected in the legend of the 1:5,000,000 tectonic map of the U. S. S. R. [11]. It was subsequently developed in more detail by geologists of the All-Union Aerogeological Trust, with this author participating [1].

Recent field work in Kuznetsk Alatau and the geological surveys on various scales in which we also participated, have yielded new and voluminous material on the tectonics of that region.

It has been determined that these structures are quite complex and diversified. Thus, some areas are characterized by large, isometric to slightly elongated, gentle anticlinal and synclinal structures complicated by a fairly complex minor folding. We shall call such structures meganticlines and megasynclines. Developed in other areas are elongated, wide zones of minor folds, extremely diversified in their form and orientation; occurring among these are highly compressed crest-like isoclinal folds as well as relatively gentle brachyfolds. Large folded structures are missing in these zones, as a rule; instead, there are crisscrossing faults (trending mostly northwest and northeast) and minor zones of warping, associated with which are deposits of a number of industrial minerals. We shall call zones similar to these "intermediate".

The tectonic map shows the arrangement of

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tektonike Kuznetskogo Alatau, (pp. 30-36).

all these structural types: meganticlines, megasynclines, outcrops of ancient formations — presumably Archaean and Proterozoic — intermediate zones, and principal faults. The megastructures were defined on the basis of formation composition and the thickness of their component rocks. In addition, the map shows the areas of thick basic Sinian, Lower-, and Middle Cambrian effusives and Middle Cambrian ultrabasic and basic intrusions.

This map shows that the intrageosynclinal troughs (North and South Minusinsk) and uplifts (Batenevsk and Shorsk), which we have identified earlier by formation analysis [5], differ not only in the formation series of their Lower Paleozoic deposits and the extent of their stratigraphic sections but also in their structural features.

A typical instance of a intrageosynclinal uplift is that of Batenevsk, which outcrops in the form of a large, isometric massif, slightly elongated sublatitudinally, its entire stratigraphic section consists of various Sinian and Cambrian carbonate facies: siliceous, bituminous, reef, dolomitic, etc. This uplift, described in another work [9], is divided into several related sublatitudinal meganticlines and megasynclines which show a well-defined step structure, with individual steps differentially uplifted and separated by large longitudinal flexures or faults. Transverse flexures are widely developed on the minor folds within each step. There are no intermediate zones in the structure of intrageosynclinal uplifts.

An essentially different formation series prevails in intrageosynclinal troughs of whi[illegible] the North Minusinsk is the one we have discussed most. They are characterized by gr[illegible] stone formations (spilite-keratophyre, green[illegible] stone-porphyrite, greenstone-schist), reef limestones, siliceous limestones, graywack[illegible] porphyrites, etc. Characteristically, the Sinian-Cambrian deposits, both as a whole a[illegible] by stratigraphic horizons, are considerably thicker in the intrageosynclinal troughs than [illegible] the Batenevsk uplift (7-8 km as against 3-4 k[illegible]

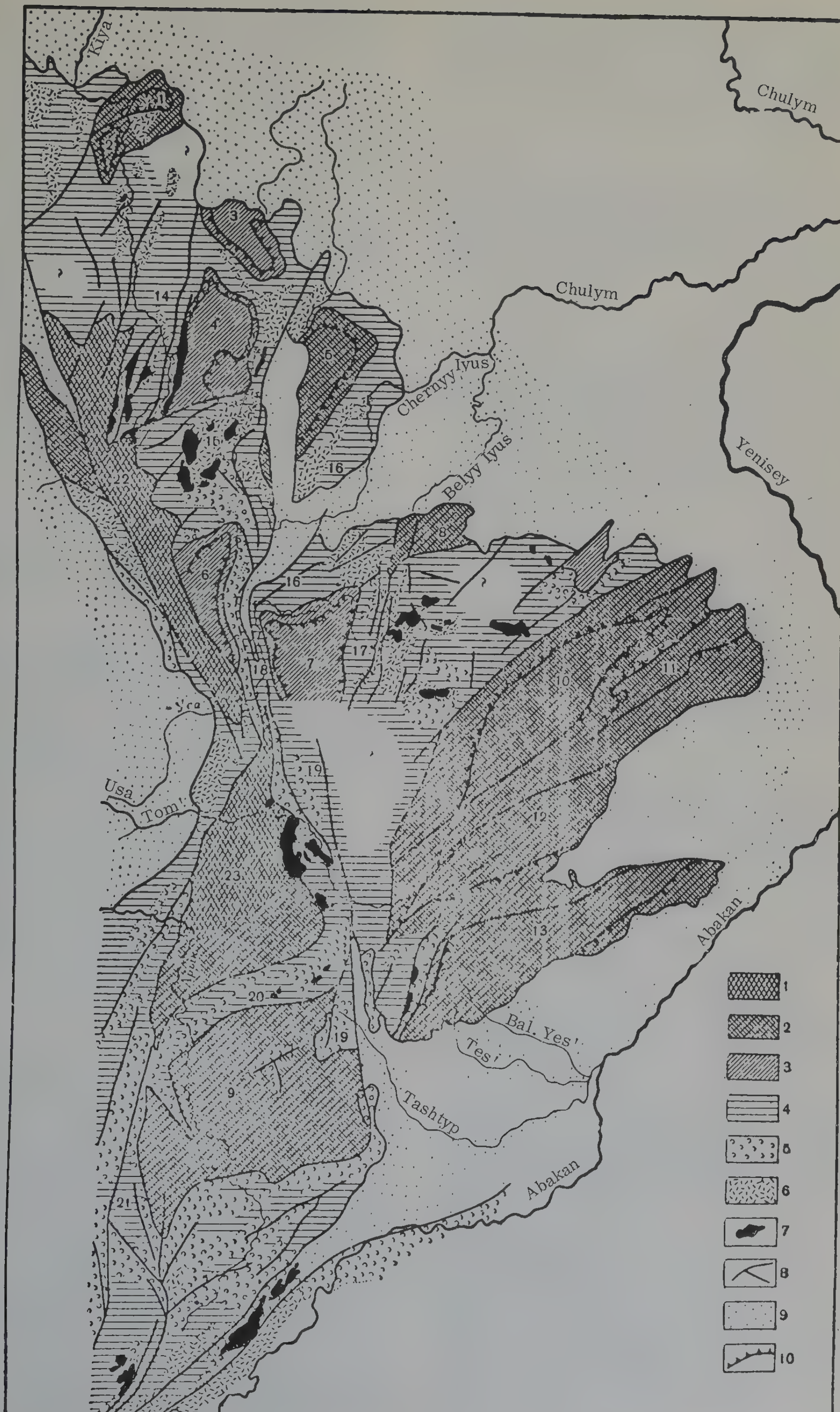
Unlike the structure of the intrageosyncli[illegible] uplifts, the structure of the intrageosynclina[illegible] troughs includes, along with isometric meganticlines and megasynclines, include extensive zones or minor folding and faulting; the latter appear to form an irregular network w[illegible] isometric megastructures at their nodes. Th[illegible] the North Minusinsk intrageosynclinal trough includes the Berikul' folded massif consisting of the Kundat meganticline and Komsomol' megasyncline; the Pervomaysk intermediate zone, the Verkne-Uryup meganticline, the Kurgusul' meganticline; the Zolotogorsk intermediate zone, the Yuzik meganticline; the Cherno-Iyus intermediate zone, the Kanym meganticline; the Usinsk intermediate zone, the East-Usinsk meganticline; the Kommunar[illegible] intermediate zone; the Yefremkinsk massif, e[illegible] In addition, presumably Archaean and Protero[illegible] zoic rocks are exposed along the southwester[illegible] and western boundary of the North Minusinsk trough, forming the northwest-trending narro[illegible] Tersinsk block and the triangular Tomsk bloc[illegible]

These meganticlines and megasynclines occur in most diverse forms varying from regul[illegible]

Tectonic map of Kuznetsk Alatau (Legend)

1 - Pre-Sinian (pre-Riphean) basement; 2 - meganticlines and megasynclines with the carbonate typ[illegible] section of Sinian and Cambrian deposits with a reduced thickness (in uplifted massifs); 3 - megant[illegible] clines and megasynclines with the carbonate-volcanic-terrigenous type of Sinian and Cambrian depos[illegible] of a greater thickness (in subsidence massifs); 4 - intermediate zones; 5 - areas with a thick mentle of basic Sinian - Lower Cambrian effusives in intermediate zones; 6 - same-for Middle Cambrian effusives; 7 - Middle Cambrian basic and ultrabasic intrusions; 8 - principal faults; 9 - Devonian and Carboniferous deposits in the Minusinsk trough of the Kuzbas and Mesozoic deposits in the West Siberian Plain; 10 - structural trends.

Numerals on the map: North Minusinsk intrageosynclinal trough (1-8, 14-18); 1 - Komsomol'sk megan[illegible] cline; 2 - Kundat meganticline; 3 - Kurgusul' meganticline; 4 - Verkne-Uryup meganticline; 5 - Yuz[illegible] meganticline; 6 - Kanym megasyncline; 7 - East Usinsk megasyncline; 8 - Yefremkinsk massif; 14 - Pervomaysk intermediate zone; 15 - Zolotogorsk intermediate zone; 16 - Cherno-Iyus intermediate zone; 17 - Kommunarovsk intermediate zone; 18 - Usinsk intermediate zone. Shorsk intrageosynclina[illegible] uplift: 9 - Batenevsk intrageosynclinal uplift (10-13): 10 - Loshchenkovo meganticline; 11 - Demidovsk meganticline; 12 - Portal'sk meganticline; 13 - Saksyr meganticline. Intermediate zones of the South Minusinsk and other intrageosynclinal troughs; 19 - Balyksinsk; 20 - Shorsk; 21 - Mrassk. Pre-Sinian basement outcrops: 22 - Tersinsk; 23 - Tomsk.


Kiya
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Chulym
Chernyy Iyus
Belyy Iyus
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ovals and triangles to isometric polygons, 10-15 to 35-40 m across. They are generally bounded by faults, or fault zones, or by steeply dipping flexures. These major structures are complicated by minor folding of low amplitude and slight pitch.

The mega-structures differ not only in form but also in the stratigraphic and lithologic features, so that it is possible to identify at least two groups of meganticlines and megasynclines by the extent of their stratigraphic section and particularly by the lithology of their Sinian and Cambrian deposits.

The first group includes mega-structures with a reduced Sinian and Cambrian section consisting largely of carbonate rocks. Such is the Yuzik megasyncline and mega-structures of the Berikul' and Yefremkinsk massifs. Almost all of their exposed section is represented by limestones, dolomites, and marbles, with poorly-developed conglomerates, sandstones, shales, and effusives — often red in color and amounting to a total thickness of 2500-3000 m. This section has a different stratigraphic range in different structures from the Sinian to the Sanashtyk horizon of Lower Cambrian Lena stage, in the Berikul' massif and the Yuzinsk meganticline; from the Sinian to the Irbinsk horizon to the Middle Cambrian Amginsk stage, in the Yefremkinsk massif. In all structures of this group, deposits younger than the carbonate interval are poorly developed.

These mega-structures are quite similar to those within intrageosynclinal uplifts from which they differ in a smaller area, and in that they are surrounded by intermediate zones having a morphology of folded structures quite different from those of the uplifts, and show evidence of intense Sinian and Cambrian volcanism. Thus, the first-group mega-structures were formed within individual, small areas of relative uplifts.

The second group includes meganticlines and megasynclines with more complete Sinian and Cambrian sections in which carbonate rocks are subordinate. Such structures are the Verkhne-Uryup and Kurgusul' meganticlines and the Kanym and East-Usinsk megasynclines. Their section covers an interval from the Sinian through the Middle Cambrian, consisting of alternating greenstone schists, reef limestones, and carbonate-terrigenous and effusive rocks. Their combined thickness is over 5500-6000 m. The second-group structures are characterized by the extensive development of the upper half of Lower Cambrian and of the Middle Cambrian. All this indicates that the second group megastructures were formed in the most downwarped parts of the intrageosynclinal troughs.

The intermediate zones are most widely developed in intrageosynclinal troughs — and the most difficult to study. They are characterized by the following four major features: 1) the above-mentioned development of minor folds of different forms, intensity, and orientation; 2) the extreme litho-facies inconsistency with a complete change in the section over a distance of several kilometers; 3) extensive development of variously oriented faults and zones of crushing; the major faults usually trend with the intermediate zones, while smaller faults form an oblique network; and 4) the thick Sinian and Cambrian basic effusives and intrusives.

In some intermediate zones, such as the Zolotogorsk and Kommunarovsk, these igneous phenomena are superimposed on one another to form a peculiar paragensis of standard and basic intrusive formations designated as the "ophiolitic complex", by V. A. Kuznetsov and G. V. Pinus [12] who associate it with deep rifts.

The range of the basic igneous activity is more restricted in age in other zones: Sinian to the base of Lower Cambrian in the Balyksinsk intermediate zone; upper half of Middle Cambrian in the Pervomaysk zone.

All this suggests that the structure and arrangement of these intermediate zones reflect the presence of deep rifts in the ancient pre-Sinian substrata, that these rifts served as channels for basic magma, and that they took place in different periods and proceeded at different rates. This is also true of meganticlines and megasynclines.

It should be noted in this connection that these intermediate zones must not be identified with the trough zones of the geosynclinal downwarps, as some investigators have done. This is because the intermediate zones, judging from their morphology and development features, were not the segments of maximum subsidence but rather the boundary zones between residual massifs which underwent major differential vertical movements.

The data cited lead us to conclude that the major structures of the Kuznetsk Alatau — the meganticlines and megasynclines — were formed on the site of residual blocks of an ancient substratum; these blocks behaved, in the Sinian and Cambrian, as isometric segments of geosynclinal troughs and relative uplifts. On the other hand, the intermediate zones separating them reflect zones of deformation in the ancient basement; as such, they are characterized by the stronger contrasts of tectonic movement and by the greater scope of basic effusive and intrusive activity. The peculiar aspect of the Kuznetsk Alatau tectonics lies in this very process of its geosynclinal development, wherein individual isometric massifs underwent great differential vertical displacements.

.t follows, then, that the distinctive tectonic ·ure of the Kuznetsk Alatau is the step-like ld-up of its mega-structures, their slight h, and the isometric angular form; they connected either by large flexures and ociated faults or by intermediate zones of nor folds and faults. According to the mation analysis data, isometric segments of synclinal troughs and relative uplifts existed his area in the Sinian and Cambrian, sur- nded by zones of vigorous basic effusive and usive volcanism. All of these facts sub- ntiate the conclusions made on the basis of re general considerations by V. V. Kho- ntovskiy [13], V. S. Meleshchenko et al. [8], V. A. Unksov [14], to the effect that the ian-Cambrian geosynclinal system in the stern Altay-Sayan area, including the Kuz- sk Alatau structures, was developed on a id, apparently sialic, basement, as it was dually broken up.

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Ministry of Geology and Mineral Conservation of the U. S. S. R.,
Moscow

Received, 22 March 1961

THE TALAS-FERGANA LATERAL DISPLACEMENT[1]

by

V. S. Burtman

This fault was described from its occur-ce in the Talas and Atoynok Ranges by \. Nikolayev [10, 11]; it was traced in the :htalyk Range by V. N. Ognev [12] and :her on by A. V. Peyve [15] and by N. M. .tsyn [19] who named it the Talas-Fergana lt.

The idea of a lateral displacement along fault was first voiced by V. N. Ognev [12, . By correlating the Upper Carboniferous ch of the Yassa River basin on the west e of the Fergana Range with similar de-its in the Maydantag Range, he arrived at splacement of 130-150 km. According to \. Nikolayev, it was approximately 75 km . L. B. Vongaz [4, 5] compared the Paleo- structural-facies zones on both sides of fault and concluded that the amount of dis-cement increases toward the northwest ining 180 km in the Central Tien-Shan. concept of the lateral nature of the Talas-gana fault was supported also by A. V. ve [16].

V. I. Popov [17], N. M. Sinitsyn [19], and e. Dovzhikov [7], and others hold a dif-ent view, to the effect that there is no izontal displacement along the Talas-gana fault and that the spatial relationships ween Paleozoic structures and facies ob-ved along it are the original ones. The essary premise of such a concept is that the as-Fergana rift was a normal fault, active ing the entire Middle and Upper Paleozoic n it divided the areas with different sedi-tary and tectonic conditions and affected primary distribution of contemporaneous es. More specifically, V. I. Popov [17] ced this idea with respect to Middle Paleo- facies along the north segment of the lt, Ye. I. Zubtsov [8], with respect to the th segment; and N. M. Sinitsyn [19] and A. Dovzhikov [7] — with regard to the entire lt.

Arguments for and against the horizontal movements along the Talas-Fergana fault were based primarily on correlations of structural-facies zones and sub-zones on the opposite sides of the fault. Despite the considerable literature on this subject [5, 17, 19], Ognev, writing a quarter of century after the first mention of a possible lateral movement along it, has recently admitted that this question is still solved [14].

Recent geologic surveying in Tien-Shan affords means of attacking this problem by comparing the general characteristics of structural and facies zones and by a comparative analysis of their inner structure, rather than by comparing isolated phenomena. This article presents the results of a litho-facies analysis of Upper and Middle Devonian deposits in the Chatkal and Naryn zones of the Central Tien-Shan and partly of the Fergana zone of the South Tien-Shan — all adjacent to the Talas-Fergana fault.

MIDDLE AND UPPER DEVONIAN LITHOLOGIC COMPLEXES

Five sedimentary groups can be identified among the Middle and Upper Devonian rocks: 1) coarse clastic red to motley conglomerates, puddingstones, gravels, and polymictic sandstones; 2) essentially quartz sandstones, mostly fine-grained, light in color, and quartzitic in aspect; 3) carbonate-terrigenous rocks — alternating sandstones, siltstones, argillites, marls, and limestones; 4) argillaceous and arenaceous limestones; and 5) limestones and dolomites free of terrigenous material. Effusive rocks are also present.

Givetian and Frasnian stages. Four types of sections have been identified corresponding to the four lithologic complexes (Figure 1). They replace each other laterally in this part of the Tien-Shan.

I. The Tayalmysh lithologic complex is essentially quartz sandstones. Its type section is located on the upper reaches of the

[1] O Talaso-Ferganskom sdvige, (pp. 37-48).

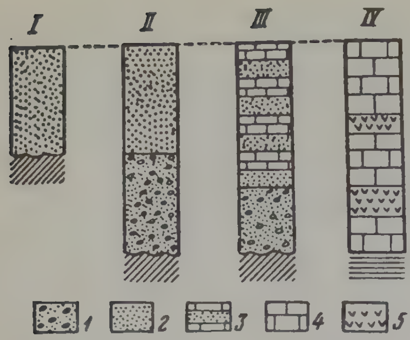

FIGURE 1. Lithologic complexes of Givetian-Frasnian deposits (I - IV; see text)

1 - Clastic rocks; 2 - essentially quartz sandstones; 3 - carbonate-terrigenous rocks; 4 - limestones and dolomites; 5 - effusive rocks.

Tayalmysh River (Figure 2) at the junction of the Pskem and Talas Alatau Ranges. It begins with a three-meter breccia bed overlain by fine-grained, essentially quartzitic-quartitoidal green-gray sandstones about 700 m thick. In the Chatkal structural-facies zone, the Tayalmysh complex area (Figure 2-B) occupies part of the Ugam, Pskem, and Sandalash Ranges, the south slope of the Talas Alatau, and the Maydantal Range.

On the opposite side of the Talas-Fergana fault, in the Naryn structural-facies zone, this complex covers a part of the Kokirimtau Mountains and the Moldotau Range. On the south slope of the Kokirimtau Mountains in the valleys of the Baydamtal (Figure 2, 42) and Kazyk (Figure 2, 42a) Rivers, Ordovician deposits are overlain unconformably by a 1000 meter formation of light-colored, essentially quartz sandstones which contain interbedded puddingstones and conglomerates up to 2 m thick. Resting conformably on these are limestones containing a Famennian fauna [9]. A similar section was described in 1957, by Ye. I. Zubtsov, in the Keninbel River valley near the Talas-Fergana fault (Figure 2, 41).

II. The lower part of the Akkapchigay lithologic complex consists of coarse-clastic rocks, and of essentially quartz sandstones in the upper section. Its type section is located in the Akkapchigay River valley on the south slope of the Pskem Range (Figure 2, 10). Reading upward it is as follows:

1. Green conglomerates and polymictic sandstones, 340 m;

2) Red polymictic sandstones and conglomerates, 290 m;

3) Light-colored quartzitic-quartzitoidal sandstones, 700 m.

Above that are limestones containing Famennian and Etrennian faunas.

There is a gradual transition from the Akkapchigay to Tayalmysh lithologic complexes from the Akkapchigay River (south branch) along the right side of the Sandalash River valley. The coarse clastic deposits gradually become thinner: along the Chon-Ishak-Ul'dy River (Figure 2, 11) the green conglomerates and sandstones are 25-30 m thick, and red sandstones and conglomerates — slightly over 100 m thick. In the glacial cirques on the right side of the Tayalmysh valley these clastics thin down to a few meters, except for the overlying quartzitic sandstones which retain their thickness. The Akkapchigay-Tayalmysh transition is accomplished by a wedging-out of lower horizons. The transitions between other lithologic complexes are effected mostly by lateral replacement.

The Akkapchigay lithologic complex is developed in two areas — the northern and the southern — located northwest and southeast of the Tayalmysh complex area, respectively. The northern area, in the Chatkal zone (Figure 2, A), occupies a portion of the north slope of the Ugam Range — Mt. Karakus — and then enters the Borolday Range. M. I. Arsovski [2] has described a section from the Kairshakhty basin (Figure 2, 3), which is quite similar to the Akkapchigay River (south branch) type section. Here, Upper Ordovician rocks are overlain unconformably by 200-1000 meters of green-gray polymictic limestones containing interbedded gravels and conglomerates. These are overlain by red polymictic sandstones up to 200 m thick. Still higher there are about 830 m of siltstones overlain by limestones containing a Famennian fauna. A similar section (Figure 2, 5) has been observed in the Dzebaglinsk Mountains [18].

In the Naryn zone the north area of the Akkapchigay complex is located in the Takhtaly Range and in the western part of the Kokirimtau Mountains. At Keninbel Pass, near the Talas-Fergana fault, the lower part of this section (Figure 2, 7) is represented by conglomerates containing interbedded sandstones (325 m). They are overlain by essentially quartzitic-quartzoidal light-colored sandstones (315 m), followed conformably by Famennian limestones [9]. Northwest along the left bank of the Kapkatash River (Figure 2, 6), the Givetian-Frasnian deposits have, according to T. A. Dodonova, a similar structure and a thickness of 1300-1500 m.

In the Chatkal zone, the boundary between

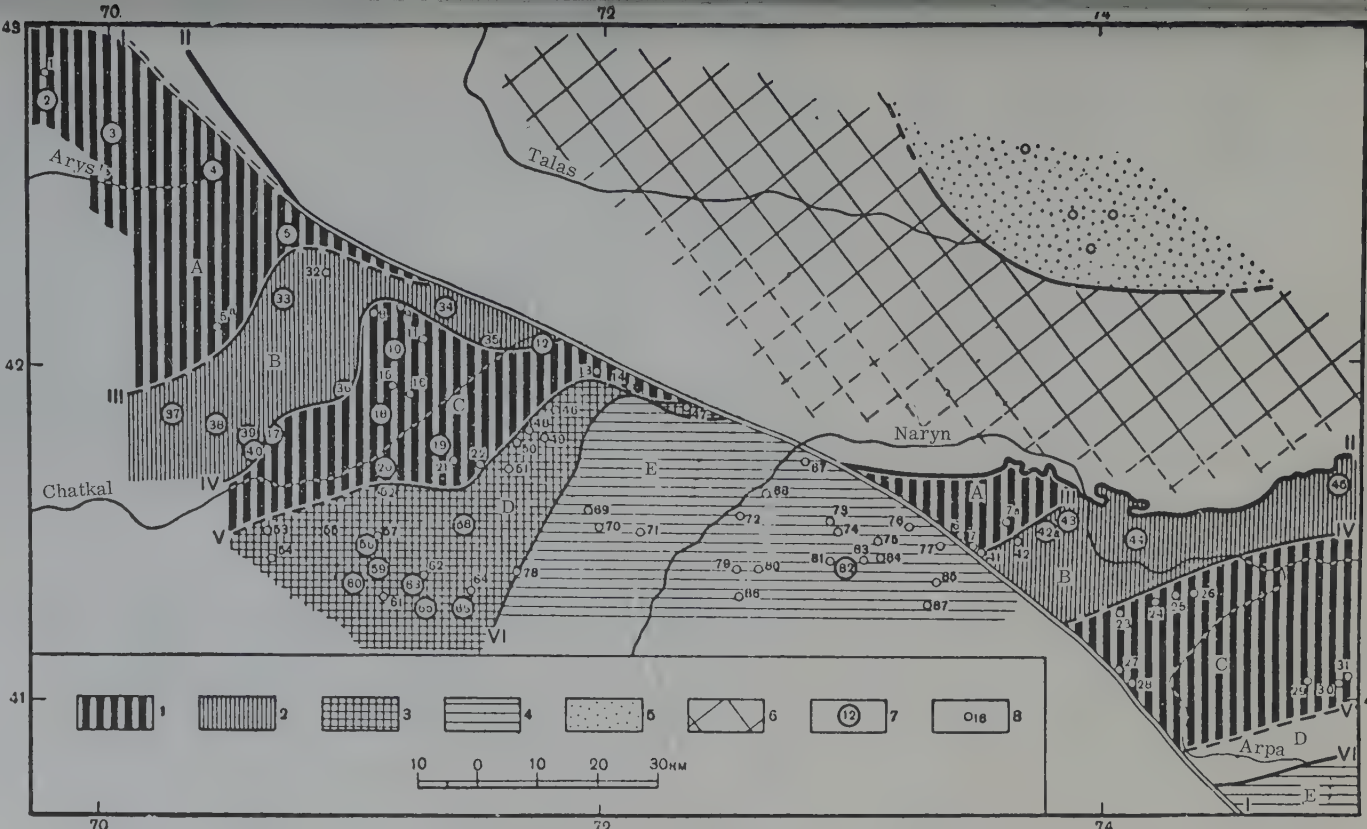

FIGURE 2. Map of the lithologic complexes of the Givetian-Frasnian deposits

Areas of lithologic complexes: 1 - Akkapchigay; 2 - Tayalmysh; 3 - Alabuk; 4 - Bozbutau; 5 - terrigenous deposits in the North Tien-Shan; 6 - area of denudation, location of stratigraphic sections: 7 - complete; 8 - incomplete; I-I - Talas-Fergana fault; II-II - main structural line of Tien-Shan (after V.A. Nikolayev). For the explanation of other numerals, see text.

the Tayalmysh lithological complex area and the north area of the Akkapchigay complex is in the Ugam Range (Figure 2, III). In the Naryn zone, it branches off from the Talas-Fergana fault at the upper Keninbel River, and extends northeastward to the Kokirimtau Mountains.

The southern part of the Akkapchigay lithologic complex, in the Chatkal zone (Figure 2 B), is located within the Pskem, Sandalash, and Chatkal Ranges; it also includes the south slope of the Talas Alatau and some of the Atoynok Range. The following section can be observed in the north slope of the Atoynok Range near the Talas-Fergana fault, in the Okum River valley (Figure 2, 14):

1) Red conglomerates interbedded with polymictic sandstones, locally tuffaceous. Visible thickness about 500 m;

2) Quartz sandstones, oligomictic, quartzitic, pink to green-grey, alternating in the lower part with calcareous argillites, 400 meters thick.

These are overlain conformably by Famennian limestones.

In the Naryn zone, the southern area of the Akkapchigay complex occurs in the Chaartash and Akshiyryak Ranges, a portion of the east slope of the Fergana Range, and extends as far as the Moldotau and Dzhaman-Davan Ranges. Givetian-Frasnian sections (Figure 2, 25 and 26) in the western part of the Akshiyryak Range were described by A. A. Luyk, in 1954. They begin with conglomerates having an exposed thickness of more than 1000 meters overlain by about 500 m of arkosic sandstones. According to the same author, exposures of the Akkapchigay complex are known to the west, in the eastern part of the Chaartash Range (Figure 2, 23) and on the east slope of the Fergana Range (Figure 2, 27). In the Dzhaman-Davan Range, the Givetian and Frasnian deposits are exposed on its northern slope (Figure 2, 29-31). The lower part of the exposed section consists of green tuffaceous conglomerates, breccias, and puddingstones, alternating with red-brown sandstones. These are followed by red sandstones and then by light-colored, fine-grained quartz sandstones.

We shall now trace the boundary between the south area of the Akkapchigay complex and the Tayalmysh complex area (Figure 2 IV). In the Chatkal zone, a transition section was described by N. V. Zhitkova in 1956 at Kokuybel Pass (Figure 2, 12). From there the boundary extends westward, then still farther west of the Karakasmak River, it veers northwest almost parallel to the Talas-Fergana fault. In the Shavursay River basin, on the northern slope of the Pskem Range, it turns south extending to the south slope of that range and is traceable in a southwesterly direction beyond the area being studied. In the Naryn zone, this bounda[ry] passes through the Toguztoraus downwarp between the Chaartash and Akshiyryak Ranges on the one side, and the Kokirimtau and Kavaktau Mountains on the other.

III. The Alabuk lithologic complex is distinguished by the presence of carbonate-terrigenous rocks. This section usually consists of a lower coarse-clastic member and an upper carbonate-terrigenous, occasionally separated by essentially quartz sandstones. Its type sections were described by N. M. Sinitsyn [1], and by L. I. Turbin, in 1955 from the Alabuk River basin (Figure 2, 58) on the south slope of the Chatkal Range. In the vicinit[y] of the Talas-Fergana fault, an incomplete section of this complex has been observed in the water-divide ridge between the Atoynok River and a right tributary of the Kol' River (Figure 2, 47). This section shows alternating limestones, essentially quartz sandstones, and argillites, all gypsiferous. Their exposed thickness is 400 meters. They are overlain conformably by Famennian limestones.

In the Chatkal zone, the Alabuk complex (Figure 2 D) is exposed southeast of the Akkapchigay complex, in the Chatkal and Atoynok Ranges (Figure 2). In the Naryn zone immediately south of the Akkapchigay complex area, Givetian-Frasnian deposits are concealed by younger deposits.

The boundary between the Akkapchigay and Atoynok complexes (Figure 2 V) extends westward along a branch of the main Talas-Fergana fault near the mouth of the Kol' River toward the Iralga valley. Farther on, this boundary runs along the Chatkal-Atoynok fault, in the Okum River basin, and then northwestward along the upper reaches of the Karasu River where it veers sharply southwestward and can be traced along the divide part of the south slope of the Chatkal Range and the lower Mynzhilka course. Farther west, the boundary trend changes to near-longitudinal; in the upper Terek course, it crosses to the north slope and apparently continues to the Ters River basin.

The southeastern boundary of the Alabuk complex (Figure 2 VI) in the Atoynok Range and eastern part of the Chatkal Range runs along the Chatkal-Atoynok fault. This fault branches off from the Talas-Fergana fault in the Atoynok-Ustasay divide and continues west-northwest over the northern slope of the Atoynok Range, toward the upper course of the Okum River. In the Okum basin it splits into several branches. The main branch which defines the Chatkal structural-facies zone in the south, continues toward the south end of Lake Sarychelek. From there, the boundary of the Alabuk complex, still on a southwesterly trend, continues underneath the Mesozoic mantle, toward the Tuyatash

ntains (Figure 2, 78) where the deposits transitional to the Bozbutau complex.

V. The Bozbutau lithologic complex cons of limestones, dolomites, and effusives. present in the South Tien-Shan area being ied (Figure 2 E). Its base, in the Bozbutau ntains, consists of acid effusives with an osed thickness of over 1000 meters (Figure 9-71). Higher in the section, the effusives nge to limestones containing the Brachiopods trypa kadzielniae Gürich [1]. According . V. Ivanov's 1948 data, these limestones about 600 m thick. They are overlain formably by Famennian limestones. East he Bozbutau Mountains the effusives are c to average. A Givetian-Frasnian section described by V. N. Ognev in 1947 south of chak Pass in the Konkol Range near the as-Fergana fault (Figure 2, 77): limestones aining an Eiffelian fauna are overlain connably by spilites whose upper section ines a 200 m limestone horizon containing phiora ramosa Phill. The exposed thickness his section is about 1000 meters. Southd, along the Kuroves River (Figure 2, 85), exposed thickness is over 4000 meters [6]. t of the Talas-Fergana fault, carbonate ks comprise the Givetian-Frasnian section he Dzhangdzhir structural-facies zone [14].

The Tayalmysh, Akkapchigay, and Alabuk plexes are transgressive over various r rocks. The Bozbutau complex is locally formably on the underlying Eiffelian deits. The Givetian-Frasnian age of these plexes is substantiated by the following . A Givetian fauna has been identified in Bozbutau complex in many areas (Figure 0, 73, 74, 77, 79, 81-83); an Upper Dean fauna has been collected from the upper ions, in the Bozbutau Mountains (Figure 2, and in the Baubashat Range (Figure 2, 82).

n the Alabuk complex, the Givetian fauna nainly Emanuella takwanensis Kaus. and ngocephalus burtini Defr. — is known from ions 56, 58, 59, 62, 63, 65 (Figure 2). nennian Theodossia schülkei Kaus., T. ssofi Vern., and Cyrthospirifer murchisonis Kon. have been collected from sections 50, 63, and 66. In most sections, the upper nday of the Frasnian is distinctly marked by appearance of Cyrthospirifer archiaci ch. and other Famennian brachiopods.

The age of the Akkapchigay and Tayalmysh nplexes is determined from their stratiphic position — conformably overlain by ks with a Lower Famennian fauna; from the nennian Bathriolepis sp. in the lower half of section of quartz sandstones in the Koksa in in the Pskem Range [1]; and from the a and fauna in the Kara-Tau Range [2]. The er boundary of the Frasnian is distinctly ked, in most instances, by the change of clastic rocks to limestones containing a Famennian fauna. The lower contact of the Givetian-Frasnian deposits is transitional within the Central Tien-Shan and is not synchronous, from place to place.

Famennian stage. The Famennian and Frasnian are conformable throughout the area being studied. The famennian is differentiated into three lithologic complexes which replace each other laterally: 1) carbonate-terrigenous rocks; 2) argillaceous and arenaceous limestones; and 3) limestones and dolomites.[2] The carbonate-terrigenous rocks, both in the Chatkal and Naryn structural-facies zones, occur in two areas — northern and southern.

We turn now to the distribution of the Famennian lithologic complexes from north to south (Figure 3).

The northern area of carbonate terrigenous rocks (Figure 3 A), takes in part of the Borolday and Ugam Ranges. In the Borolday Range along the upper Kairshakhta course (Figure 3, 1), the Famennian section consists of silty and sandy limestones alternating with silts, organoclastic limestones and rare intercalations of polymictic sandstones. The limestones contain Camarotoechia cf. turanica Rom. The total thickness is about 300 m [2]. In the Naryn zone, this area includes segments of the Takhtalyk Range and the Kokirimtau and Kavaktau Mountains. Famennian deposits described by V.I. Nasredinov and T. A. Dodonova along the Keninbel' River (Figure 3, 8) near the Talas-Fergana fault are represented by limestones alternating with calcareous siltstones and occasional silt intercalations. The limestones contain Cyrthospirifer cf. archiaci Mürch. and C. aff. brodi Wen. Total thickness, 350 meters.

The argillaceous and arenaceous limestones in the Chatkal zone (Figure 3, B) occur in the Pskem and Maydantal Ranges and partly in the Borolday, Ugam, and Chatkal Ranges. In the Borolday Range, in a river valley of the same name, the Famennian section consists of argillaceous limestones and marls containg Cyrthospirifer romanowskiy Nal. and C. brodi Wen. (Figure 3, 38). Its thickness is 1300 m [2]. As observed by the author, quartzitic sandstones in the northeast part of the Pskem Range are overlain by dark-grey, slaty, biodetrital argillaceous limestones up to 100 m thick. Collected from this horizon in the Akkapchigay (north branch)-Tastarsay watershed (Figure 3, 41) were Famennian brachiopods,

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[2]Each Famennian lithologic complex consists of a single rock group, which obviates the necessity of naming them by type localities.

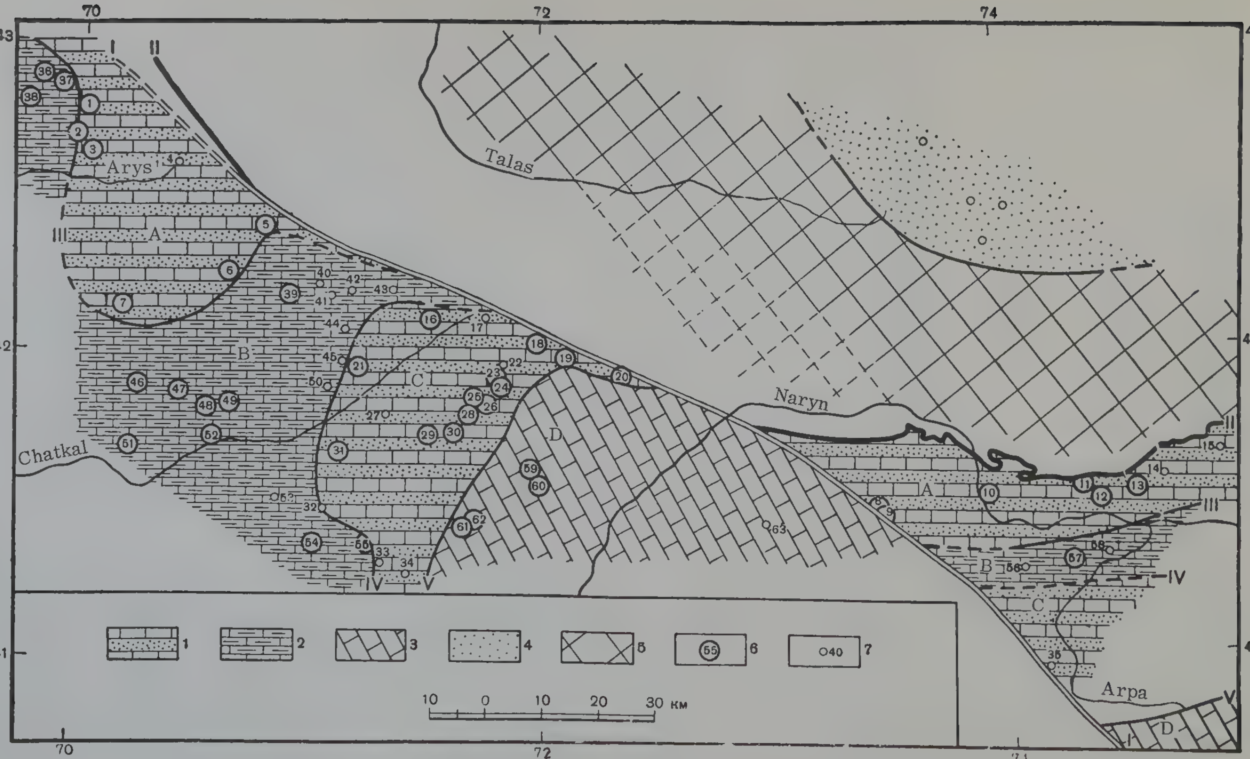

FIGURE 3. Map of Famennian lithologic complexes

Distribution areas: 1 - carbonate-terrigenous rocks; 2 - argillaceous and arenaceous limestones; 3 - limestones and dolomites; 4 - terrigenous deposits in North Tien-Shan; 5 - area of denudation. Location of stratigraphic sections: 6 - complete; 7 - incomplete; I-I - Talas-Fergana fault; II-II - main structural line of Tien-Shan (after V.A. Nikolayev). For the explanation of other numerals, see text.

hospirifer aff. archiaci Mürch., C. cal- tus Sow., C. sulcifer H. C., and Hypo- is acmolla Nal., typical of higher horizons. Adelung [1] and L. I. Turbin in 1958 ob- ed intercalations of calcareous sandstones e argillaceous and arenaceous limestones Cyrthospirifer archiaci Mürch, in the beltesh valley in the Sandalash Range ure 3, 45). This is a section transitional e carbonate-terrigenous complex. The ennian section (Figure 3, 5) in the baglinsk Mountains [18] also appears to ansitional.

the Naryn zone, this complex occurs in Chaartash and Aksiyryak Ranges. Accord- o A. A. Luyk (1955), the lower Famennian ions are limestones, partly sandy, overlain haly limestones (Figure 2, 56-58). Numer- Famennian brachiopods were collected e. The boundary between these argillaceous arenaceous limestones (Figure 3 III) and the hern carbonate-terrigenous area in the lday Range extends southward, then turns , crosses the Ugam Range and approches anite intrusion at the Talas-Fergana fault. e Naryn zone, this boundary lies in the Toguz- u depression north of the Chaartash and iiyryak Ranges.

he south carbonate-terrigenous area (Fig- 3, C) in the Chatkal zone, occupies the ern part of the Chatkal Range, part of the lalash Range, the southern slope of the s Alatau and the northern slope of the nok Range. The following section of nennian deposits is exposed in the Okum ey (Atoynok Range) near the Talas-Fergana (Figure 3, 19):

) Argillaceous and arenaceous limestones aining intercalations of essentially quartz lstones, 40 meters.

) Alternating limestones and calcareous stones, 120 m.

. N. Gavrilova identified Cyrthospirifer li Ven. and Spirifer aperturatus Schlot, n the upper part of this section.

amennian outcrops are known from the yn zone south of the Akshiyryak Range, in Pshan River valley near the Talas-Fergana t (Figure 3, 35). According to Ye. I. tsov [6], they are represented by alterna- calcareous siltstones, argillites, and llaceous limestones containing Cyrtho- ifer aquilinus Rom. and Camarotoechia nica Rom. The exposed thickness is 175 ers.

The boundary between the southern car- ate-terrigenous area and the argillaceous arenaceous limestones (Figure 3, IV) in Chatkal zone, runs from the middle Sumsar River to the Sumsar-Kassansay divide in the north, then northwestward, toward the upper course of the Kassansay. From there it continues north-northeast to the Kumbel' River basin in the Sandalash Range and farther north-east along the water-divide part of that range to the Talas-Alatau. In the Naryn zone this boundary runs south of the Akshiyryak Range within the Naryn depression. The southeastern boundary of these carbonate-terrigenous rocks (Figure 3, V) runs northeastward between the Bozbutau Mountains and the Chatkal Range and then along the Chatkal-Atoynok fault, which is the south boundary of the Chatkal structural-facies zone.

The limestones and dolomites (Figure 3, D) constitute a lithologic complex developed in the Bozbutau Mountains (Figure 3, 59-62) and in the Baubashata Range (Figure 3, 63), west of the Talas-Fergana fault. A Famennian fauna has been identified in the limestones of both areas. East of the Talas-Fergana fault, Famennian limestones and dolomites are developed in the Dzhangdzhir structural-facies zone [14].

THE NATURE OF THE TALAS-FERGANA FAULT

I. There are identical lithologic complexes on both sides of the Talas-Fergana fault. Their respective development zones, in contact with the fault, continue on its opposite side, with a considerable offset. The west side has been relatively displaced to the northwest (right lateral shift).

• Thus, the Talas-Fergana fault crosses the development areas of several lithologic complexes, rather than constituting a boundary between them; i. e., it does not affect their primary distribution. Similar conclusions on the absence of the effect of this fault on Middle Paleozoic facies were drawn from various data by V. A. Nikolayev [6] and L. B. Vongaz [4] for the Fergana segment of the fault, and by this author [3], for its northern segment. Material presented in this article corroborates these conclusions and disproves the concept of N. M. Sinitsyn [19], A. Ye. Dovzhikov [7], and others.

II. A thick section of limestones and basic lavas of the Bozbutau lithologic complex occurs in the water-divide portion of the Kenkol Range between the Sarytash headwaters and the right tributaries of the Kokirim River, on the west side of the Talas-Fergana fault (Figure 3, 77, 85), while thick conglomerates and sandstones of the Akkapchigay and Tayalmash complexes are present on its east side (Figure 2, 6, 7, 41). A fault contact of obviously different but contemporaneous facies, in the absence of vertical movements, is evidence of a lateral displacement.

III. The boundaries of lithologic complexes abut at the Talas-Fergana fault. In the Chatkal zone, the south boundary of the Tayalmysh complex (Figure 2, IV), the Akkapchigay-Alabuk complex boundary (Figure 2, V), and the south boundary of the Alabuk complex (Figure 2, VI), change their northeasterly trend to the southeast, 15-20 km short of the fault and extend for some distance almost parallel to it before finally reaching it. This bend of the faults and the axes of the folded structures in the vicinity of the Talas-Fergana faults is common in this region. It should be emphasized that the two boundaries of the Tayalmysh complex (Figure 2, III, IV) and to some extent, the Akkapchigay-Alabuk boundary (Figure 2, V), are not associated with faults.

Thus, the boundaries between the Middle Devonian lithologic complexes, as well as the trend of structures and of the faults cutting them, exhibit a bend parallel to the Talas-Fergana fault in its vicinity; this suggests an epigenetic origin of the bend.[3]

This phenomenon is an instance of deformation of a peculiar geologic body made up of rocks of a definite lithologic complex. The boundaries between these lithologic complexes outline a fault-controlled fold with the crest facing north-northwest. The presence of this fold corroborates the lateral nature of the displacement along the Talas-Fergana fault, while the orientation of the fold's axis with respect to the fault trend indicates a right lateral shift.

Data are inadequate on the deformation of lithologic complexes in the Naryn zone. However, the presence of such phenomena can be inferred because here, too, there is a bend of folds and faults in the vicinity of the Talas-Fergana fault (facing south, as anticipated). Here, however, the extent of this bend is not as great as in the Chatkal region.

The origin of the fault-controlled fold is associated with a plastic[4] displacement of material in the fault sides. In determining the magnitude of this displacement in the Chatkal zone, we project the boundary of the lithologic complexes from where they begin to bend and outline the crest part of the fault-controlled fold (Figure 4). The distance P_1-P_3 between the projected intersection with the fault and the true point of intersection of the fault and the boundary is regarded as the distance of lateral displacement. It is 50-60 km, as dete[r]mined from the south boundary of the Tayalys[h] complex of the Givetian-Frasnian deposits, a[nd] about 40 km, from the Akkapchigay-Alabuk boundary. The position of the southern Chatkal boundary also, suggests a plastic displacement of about 40 km (Figure 4). A round[ed] figure for the Chatkal zone is about 50 km (P_1, Figure 4).

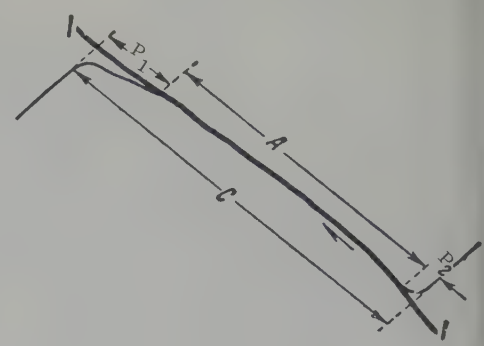

FIGURE 4. Diagram of the Talas-Fergana fault (I-I)

See explanation in text

IV. The apparent displacement (Figure 2, A) can be determined by measuring the distanc[e] between the boundaries of the corresponding complexes touching the fault in the Chatkal and Naryn zones. The total displacement (Figure 4, C) is the sum of the apparent and the plastic displacement. We shall attempt to compute it.

The best exposure in the Naryn zone is the intersection of the Talas-Fergana fault and the northern boundary of the Tayalmysh complex i[n] the upper course of the Keninbel', a triburary o[f] the Kokirim. In the Chatkal zone, this bounda[ry] passes along the Ugam water-divide. Igneous rocks, present here in the vicinity of the fault, make it impossible to trace the fault-controlle[d] fold. The total displacement can be determine[d] from the relative position of the northern Tayalmysh complex boundary (Figure 2, III-III) in the Chatkal and Naryn zones. It is 250 km. Subtracting 50 km — the plastic displacement in the Chatkal side of the fault (and disregarding it on the Naryn side) — we obtain a figure of about 200 km. It comes to the same amount as determined from the southern Tayalmysh boundary (Figure 2, IV-IV), with the tota[l] displacement of 250 km. The same figure is o[b]tained by comparing the positions of the Famennian argillaceous and arenaceous limestones o[n] both sides of the fault (Figure 3, B). However, the intersection points of their boundaries with the Talas-Fergana faults are not as definite.

[3]Another explanation of this bending of the lithologic-complex boundaries as caused by the shape of the sedimentary regions must be rejected as not explaining the bend of the folds and faults.

[4]"Plastic" only with relation to the Talas-Fergana lateral fault as a whole. This displacement includes those along associated smaller faults.

As mentioned before, V. A. Nikolayev com-
ed the apparent displacement as 75 km [6]. error of that computation[5] lies in the ong correlation of Upper Devonian deposits he Atoynok Range and Kokirimtau Moun- ns. Using the terminology of this article, an be stated that he correlated dissimilar as of the same type deposits: the south a of the Akkapchigay complex in the Chatkal e (Atoynok Range) with the north area of s complex in the Naryn zone (Kokirimtau untains); also the southern area of the mennian argillaceous and arenaceous lime- nes in the Chatkal zone with the northern a of these rocks in the Naryn zone (Figures nd 3).

In conclusion, it should be added that the uth boundaries of the Naryn and Chatkal uctural-facies zones — the Abtash [8] and atkal-Atoynok [1] faults — also reach the las-Fergana fault line at a distance of about 0 km — the apparent lateral displacement ong it in the Central Tien-Shan.

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Received, 24 November 1960

[T]HE FEASIBILITY OF LONG DISTANCE, HORIZON-BY-HORIZON CORRELATION OF FLYSCH SECTIONS ("TELECONNECTION")[1]

by

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The method for the layer-by-layer correla-[tio]n of flysch sections was worked out by N. B. [Va]ssoyevich, and M. K. Kalinko in 1938. In [th]e following period of over 20 years, this [me]thod has become popular in the U. S. S. R. [in] studying areas of flysch deposits.

N. B. Vassoyevich cites many instance of [the] detailed correlation of flysch in his special [mo]nographs [2, 4], mostly on Upper Carbon-[ife]rous carbonate flysch from various regions [of t]he Caucasus. In the Southwestern Cau-[ca]sus, N. B. Vassoyevich, M. K. Kalinko, V. V. [Ti]khomirov, and this author studied and corre-[lat]ed in detail mainly the Turonian deposits [(Ke]mch formation), as well as the Campanian [fly]sch of the Dibrara and Vandam zones and of [the] Baskal mantle [1-5]. In Gornaya Kakhetia, [N.] B. Vassoyevich correlated the Eshmakis-[h]evi, Dzhorchi, Sabuin, Mekvadur, Kviter [(Up]per Cretaceous), etc. formations [2]. In [the] Northwestern Caucasus he correlated in [det]ail, by the connection[2] method, the Genniokh [for]mation sections (Santonian) in cement plant [qu]arries, which has led to the development of [a s]ingle nomenclature for "naturals". We have [su]cceeded in correlating a number of the [Ke]rket formation sections (Turonian), and [se]veral sections of the Goryachiy Klyuch for-[m]ation, in cooperation with G. M. Aladatov. [...] L. Afanas'yev has recently done much work [in] a detailed study of the Upper Cretaceous [se]ctions in the Northwestern Caucasus.

This brief review shows the great extent of [wo]rk done on correlation by the connection [m]ethod.

It should be emphasized that while the early [co]rrelation was done on sections hundreds of meters apart (Campanian flysch at Sovietabad, [2]), and individual sections 3 km apart at the most, the subsequent distances were considerably longer. For instance, in studying the Southeastern Caucasus Turonian flysch basin (lower part of the Kemch formation), N. B. Vassoyevich and the author surveyed 40 sections in detail. The distance between the outermost section, along the trend of the flysch trough was about 130 km, with distances between individual sections — from 5 to 35 km. The correlation distance across the strike was 30-40 km, with the sections 5-15 km apart [5]. The results turned out to be satisfactory. It was possible to identify and index a member, in almost all sections surveyed and to analyze it to determine the contemporaneous paleogeographic conditions prevailing in the flysch basin.

Recently we surveyed in the Northwestern Caucasus 16 detailed sections near the base of the Kerket formation (correlative with the lower part of the Kemch formation). These sections are located along and across the trend of the Novorossiysk flysch trough. The strike distance between the outermost sections (hamlet of Gornyy and Mzymta River) was 235 air km; a transverse belt about 20 km wide, was also surveyed. It has been established that the Kerbet formation is represented by three facies: flysch, sub-flysch, and coarse flysch. The flysch sections (orthoflysch of N. B. Vassoyevich) best lend themselves to correlation; sub-flysch sections are not as easily correlated but are still quite satisfactory because of their typical marker horizons containing one er (rhythm element) of greater thickness; we had difficulty in correlating the flysch and sub-flysch sections by means of rhythmograms and we used the sections of an intermediate zone, between the areas of typical flysch and sub-flysch. The connection method did not work on the coarse flysch because of the great number of micro-erosions and the great increase in thickness.

At the same time, this author in cooperation with G. M. Aladatov succeeded in making the first correlation by the graphic comparison

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[1] O vozmozhnosti posloynogo sopostavleniya [r]azrezov flisha na bol'shikh rasstoyaniyakh (tele-[ko]neksii), (pp. 49-57).
[2] The term, "connection", for a detailed correla-[tio]n of sections by graphs was introduced by De [Ge]er.

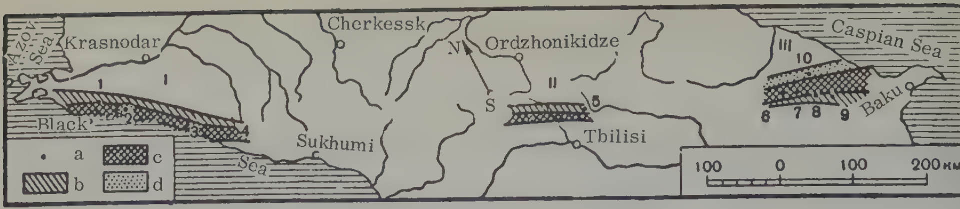

FIGURE 1. Areas of Upper Turonian flysch

I - Northwestern Caucasus; II - Eastern Georgia; III - Southeastern Caucasus; a - sections mentioned in text; facies zones; b - sub-flysch; c - flysch; d - coarse flysch. Figures in map: 1 - section along Idoumes R. (I); 2 - Kamenistaya Shchel' (K); 3 - Tsuskavadzhe R. (Ts); 4 - Mzymta R. (Chvezhipse) (Ch); 5 - Orvili R. (O); 6 - Gerdymanchay R. (G); 7 - Uzundzha R. (T); 9 - Kyzyl-Chay R. (Ch); 10 - Ariskyunsh-Chay R. (Sh).

method, of three terrigenous flysch sections (Upper Paleocene) over a total distance of 25 km (10 and 15 km).

Thus, a horizon-by-horizon correlation of flysch section was achieved over a distance of more than 200 km at 5-25 km intervals.

The question is whether we have achieved the ultimate in correlation or is it possible to correlate in detail flysch sections 500-1000 km apart. On the basis of the results achieved, it appears that the answer is in the affirmative. Our reasons are as follows.

First, the Greater Caucasus flysch zone, extending from the southeast to the northwest, through Gornaya Kakhetia, Mtiuletia, and South Osetia, is marked by the great similarity of its Cretaceous deposits. Some horizons maintain practically the same facies content all the way from the Caspian shore to Novorossiysk; such is the Ananur formation (and its stratigraphic equivalent — the Zarat formation) which is an Upper Cretaceous marker horizon [1-3]. I. M. Gubkin, in his time, noted after his work in the Anapa area, the great similarity in the Upper Cretaceous section of both regions (Il'khidag formation and the Durso series, as now understood — [6]).

Second, it can be assumed from the oscillation hypothesis for the origin of flysch that these movements, during certain periods, were very similar if not identical, throughout the entire flysch zone.

However, while the first premise requires no proof, the second needs verification, because there is no current unanimity of opinion as to the origin of flysch rhythms. Some students go as far as to deny the existence of any such rhythm [7].

In using the method of previous work, sections of a datum horizon would have to be surveyed every few kilometers to every few tens of kilometers, thus increasing the distance between the outermost points. This, however, is impossible because the Greater Caucasus flysch zone is not exposed throughout its entire length, the southeastern and east Georgian areas of Upper Cretaceous flysch being covered by younger sediments of the Alazan-Agrichay intermontane trough, while the east Georgian and northwestern areas are separated by major thrusts.

Thus, the only remaining possibility was to proceed with a detailed survey of sections separated by many hundreds of kilometers. The first such attempt was made on Upper Turonian flysch, selected for the following considerations:

1. These deposits are extensively developed in each of the three areas of the Upper Cretaceous flysch; because of the complex folding, they can be studied both along and across the strike of the tectonic and facies zones.

2. At the base of Upper Turonian formations with different names in different areas, there is a marked horizon standing out lithologically in the Upper Cretaceous section in all areas (Ananur horizon in Northwestern Caucasus and East Georgia, Zarat horizon in the Dibrara zone of Southeastern Caucasus). Its presence is an important additional correlation criterion.

3. The Greater Caucasian Upper Turonian flysch zone is rhythmically stratified and may be regarded as a typical representative of carbonate flysch.

4. Tens of Upper Turonian sections have already been surveyed in Northwestern and Southeastern Caucasus; this considerably facilitated further work.

To check up on the accuracy of detailed correlation over such a long distance (the westernmost section on the south slope of Kovdak Mountain, Southeastern Caucasus, is 750 km away from the eastermost section on the Mzymta River not far from the mouth of the Chvezhipse River, Northwestern Caucasus), it was imperative to survey at least

e intermediate section in eastern Georgia. ring our short visit in Gornaya Kakhetia, succeeded in surveying such a section of e Margalitis-Klde formation on the Orvili ver at the village of Akhmet. After that, it s possible to take up the study of the sibility of flysch correlation.[3]

As we have already mentioned, the Ananur iceous flysch formation of the Northwestern ucasus is overlain by the Kerket formation ose lower part we have studied in detail. It represented by a rhythmic alternation of ddingstones, sandstones, siltstones, limeones, marls, and marly shales.

I-a rhythm sub-element (of a polystratum)[4] usually represented by puddingstones and arse-grained sandstones, less commonly by e conglomerates (in the coarse flysch zone).

I-b per is most often represented by silty omogeneous or clastic) limestones, stratied to cross-bedded, and by less common lcareous silts and silty pelites.

II-a per is represented by light-grey to pink mestones containing some silica.

II-b per consists of grey to pink marls and een-grey to red calcareous shales.

In addition, there are occasional layers of uller's earth.

The transition from the Ananur to Kerket rmation is ordinarily gradual, except for the arse flysch zone and the transitional Kerketrmation at its base, where conglomerates e present.

In Gornaya Kakhetia and Mtiuletia, the nanur siliceous formation of the Chiaur zone lso within the Chinchvelt mantle) is overlain y the Margalitis-Klde formation of puddingones, sandstones, siltstones, limestones, nd marly shales.

The composition of base of the Margalitislde and Kerket formations is quite similar.

I-a per is represented by puddingstones, nd coarse-grained sandstones.

I-b per consists of silty limestones and alcareous silts, horizontally- to cross-stratied.

II-a per consists of greenish to pink limestones, slightly siliceous.

II-b per is represented by light-green to pink and red marls and calcareous shales.

The similarity is sometimes observed even to specific details. For instance, the II-a per is characterized by the presence of flat siliceous lenses in the Northwestern Caucasus as well as in eastern Georgia; the intensity of the red and pink hues decreases from south to north, etc. The southern sections in both regions are more of a true flysch, changing to sub-flysch northward (toward the central uplift).

Finally, the Zarat formation in the Southeastern Caucasus (corresponding to the lower and middle parts of the Ananur formation) is overlain by the Kemchi formation of rhythmically alternating puddingstones, sandstones, siltstones, limy marls, and shales.

I-a per — puddingstones, and coarse-grained sandstones with a calcareous cement;

I-b per — calcareous sandstones and (or) siltstones, horizontally- to cross-bedded and slaty;

II-a per — light-colored pelitic limestones;

II-b per — green-grey marls and marly shales. In addition, there are beds of noncalcareous shales (III er) and Fuller's earth.

Thus, all three areas show great similarity in the rock components of the Upper Cretaceous interval being studied. This similarity, even in minor details, is particularly striking in a consecutive inspections of the sections.

Figures 2, 3, and 4 are rhythmograms of the Upper Turonian carbonate flysch grouped by facies zones. It is obvious that the overall configuration of curves for sections of the same facies zone is quite similar despite their being hundred of kilometers apart (over 1000 km, in some instances), while the proximate sections (a few tens of kilometers apart) of other facies zones have a quite different aspect.

Some segments of these rhythmograms, particularly those for the flysch and sub-flysch zones, are quite similar (it should be kept in mind that in this instance we have regarded as sub-flysch the sections of a zone transitional from flysch to sub-flysch; this is the metaflysch zone of N. B. Vassoyevich).

If these sections were located in the same flysch province, we undoubtedly would have considered them as synchronous — especially those along the Mzymta and Orvili Rivers. However, such an assumption is not warranted for the time being, and we only note the great similarity of these sections.

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[3] Horizon-by-horizon correlation of sections over onsiderable distance by the graphic method, has een called "teleconnection", by N. B. Vassoyevich.
[4] Following the usage adopted in the literature on ysch, an element of rhythm will be designated as r; a sub-element of rhythm, as per [p stands for od: Russian for sub].

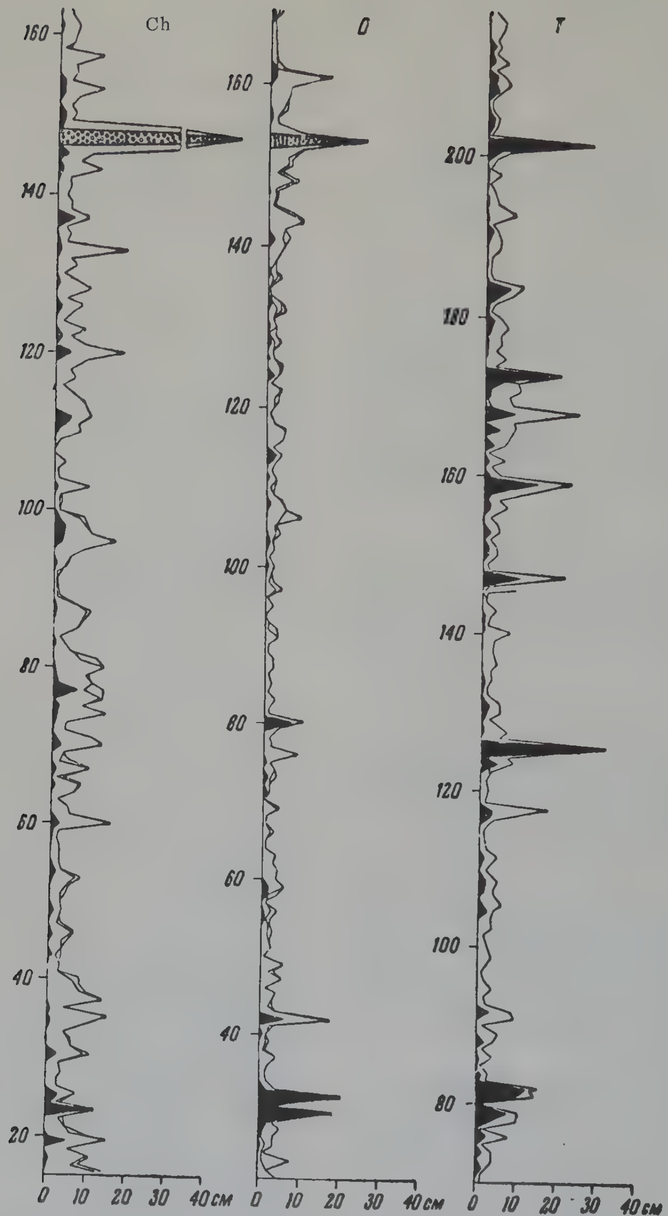

FIGURE 2. Rhythmograms of various sub-flysch sections

Ch - Mzymta R.; O - Orvili R.; T - Tuk R. Distance between Ch and O is 450 km; between O and T - 350 km.

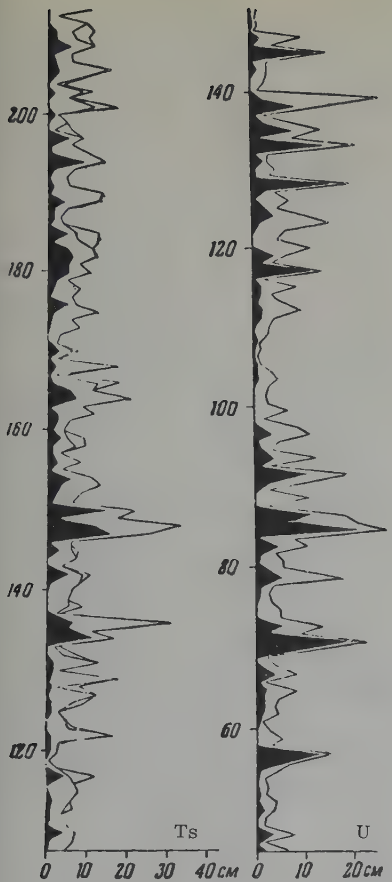

FIGURE 3. Rhythmograms of flysch sections

Ts - Tsuskavadzhe R.; U - Uzundzha R. Distance between the two, 850 km.

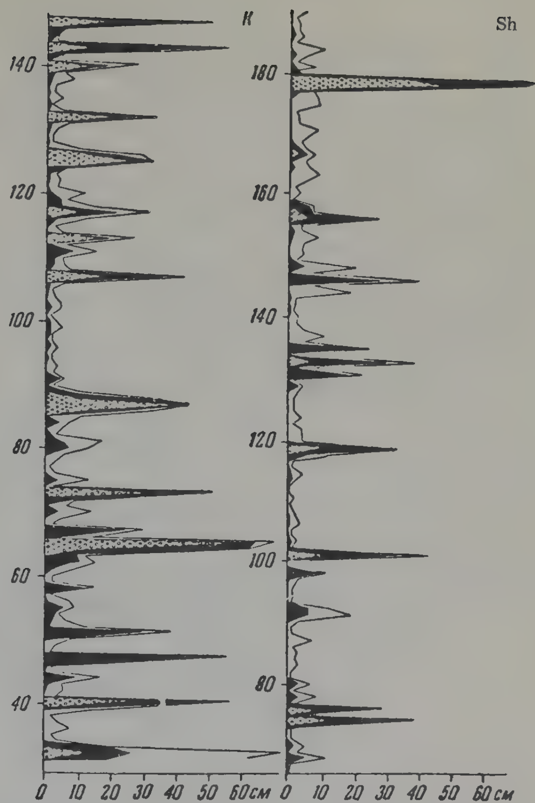

FIGURE 4. Rhythmograms of coarse flysch sections

K - Kamenistaya Shchel'; Sh - Ariskyunsh-Chay R. Distance between the two, 950 km.

As is true for individual areas, the coarse lysch areas are the least individualized. This naturally follows from the presence of numerous micro-crosions, causing many rhythms to be obliterated, while others show greater thicknesses.

The correlation can be made not from rhythmograms alone. The table below gives some data on the thickness of rhythms, their elements, and sub-elements. Except for the average thicknesses, all figures are percent values, because the members correlated are not quite synchronous. The computations have been made for an interval of 150-200 rhythms (in different sections). Data on III er and Fuller's earth are omitted (therefore, the total does not always amount to 100%).

A study of this table, as well as of the rhythmograms, reveals that sections hundreds of kilometers apart but belonging to the same facies zone are much more alike than those only 10-20 km apart but from different facies zones. This fact is of great general importance.

Certain differences are present in sections from different areas. The table shows that the role of I er and II-b per increases on the whole from west to east, while that of II-a per decreases. This is true even for the Northeastern Caucasus, taken by itself. Both N. B. Vassoyevich and myself came to the conclusion that this is due to facies transitions from II-a to II-b pers [2, 5]. Indeed, the figures for II er are quite similar, on the whole.

The other characteristics of flysch sections are best seen from the rhythms' thicknesses — by intervals, this time. Figure 5 shows the distribution of rhythm thicknesses within

Rhythm characteristics in flysch rocks

Characteristic indexes	Coarse flysch		Flysch		Sub-flysch			
	Kamenistaya Shchel' (K)	Ariskyush (Sh)	Tsuskvadzhe north (Ts)	Gerdyman-chay R. (G)	Idoumes R. (I)	Mzymt R. (M)	Orvili R. (O)	Kyzyl-Chay R. (K)
Total thickness of I-a in % of overall thickness	30.0	30.4	—	—	—	8.5	2.9	7.1
Number of rhythms with I-a, %	15.4	8.7	—	—	—	0.9	0.5	9.2
Total thickness of I-b in % of overall thickness	23.2	28.1	19.2	29.7	18.9	9.6	14.2	11.2
Number of rhythms with I-b, %	98.2	62.5	98.3	71.9	100.0	99.1	97.8	67.8
Total thickness of I er in % of overall thickness	53.2	58.5	19.2	29.7	18.9	18.1	17.1	18.3
Number of rhythms with I er, %	99.1	71.2	98.3	71.9	100.0	99.1	97.8	77.0
Total thickness of II-a in % of overall thickness	37.2	31.1	66.6	56.3	80.5	76.9	65.2	31.3
Number of rhythms with II-a, %	76.0	89.1	94.6	95.3	100.0	100.0	93.2	69.0
Total thickness of II-b in % of overall thickness	9.6	9.8	14.2	12.5	0.6	5.0	15.6	46.2
Number of rhythms with II-b, %	24.8	52.1	37.3	74.7	0.1	21.4	37.6	62.9
Total thickness of II er in % of overall thickness	46.8	40.9	80.8	68.8	81.1	81.9	80.8	77.5
Number of rhythms with II er, %	91.5	91.2	99.4	98.9	100.0	100.0	99.6	97.6
Average thickness of I-b	2.4	3.2	1.8	2.9	0.7	0.6	0.5	1.4
" " " II-a	5.0	2.5	6.5	4.2	2.9	4.5	2.5	3.9
" " " II-b	4.0	1.3	3.5	1.2	0.4	1.4	1.5	6.5
" " " a rhythm	10.3	8.1	9.3	7.2	3.6	5.9	3.5	8.4

Turonian flysch sections by logarithmic intervals (which gives a better picture than the equal intervals).

Figure 5 shows that the graphs of proximate sections from different facies zones are less similar than those of the same facies type sections, however distant apart. For instance, the graph of the Tsuskvadzhe River section (Northwestern Caucasian flysch zone) is quite similar to that for the Mt. Uzundzha section (Southeastern Caucasian flysch zone), although they are 850 km apart; on the other hand, the graph of the Tsuskvadzhe River section is quite unlike that of the Idoumes River section (Northwestern Caucasian sub-flysch section), 110 km away. There are many more such instances.

A similar picture is obtained by comparing the rhythm types from sections in various zones, as well as other indexes.

What conclusions are to be drawn from data cited?

Although we are far from being able definitely to identify individual rhythms over distances of hundreds of kilometers, it must be granted that these rhythmograms, within a facies zone, show a great similarity which cannot be a matter of chance.

The same similarity transpires from the study of data on the thickness of rhythms and of parts of rhythms, taken as an average or by intervals; here, too, the similarity within a facies zone, regardless of the distance, is greater than for proximate sections belonging to different facies zones.

These facts are readily explained on the basis of the oscillation hypothesis by postulating the same oscillatory rhythm for the entire flysch trough. This would explain the structural difference of different facies zones.

A better way to state the problem is that this distribution of thicknesses corroborates the oscillation hypothesis and refutes attempts to explain the presence of sand intercalations as being due to the action of turbidity currents. It is beyond the scope of this article to take up this subject in detail. We only state that such currents are not likely to form sequences that similar in thickness over areas 1100 km long. It is more likely that they were deposited by the long-term action of bottom currents which spread terrigenous material throughout the basin, as witness the textures of the initial

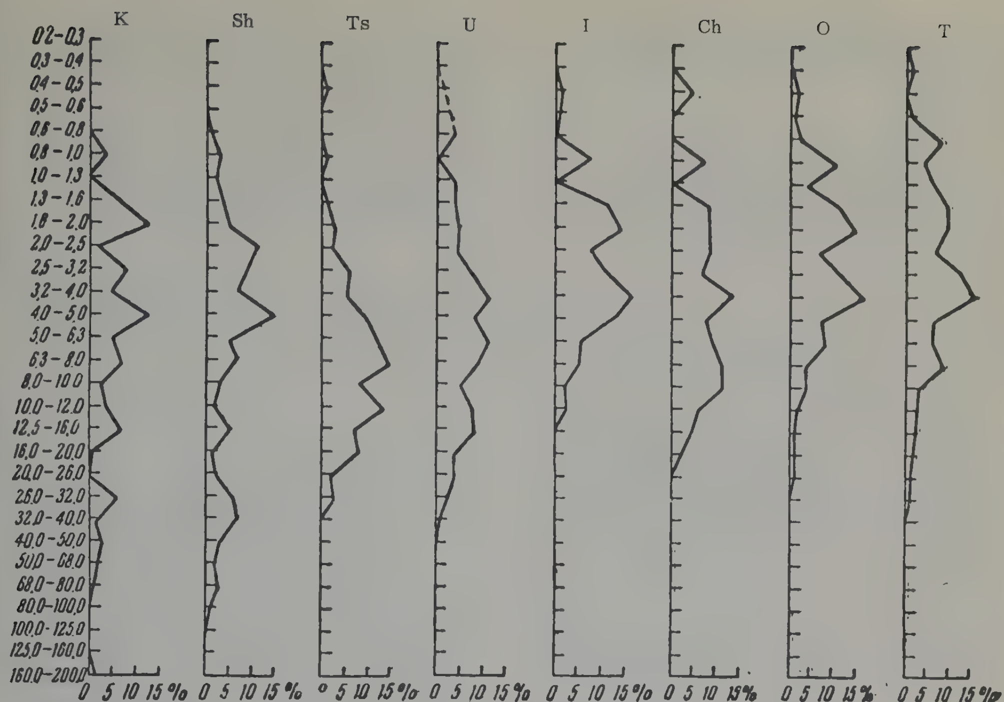

FIGURE 5. Distribution of thicknesses by intervals in a number of Turonian flysch sections

Coarse flysch: K - Kamenistaya Shchel'; Sh - Ariskyunsh-Chay R.; Ts - Tsuskvadzhe R.; U - Uzundzha R. Sub-flysch: I - Idoumes R.; Ch - Mzymta R.; O - Orvili R.; T - Tuk R.

rhythm elements (polystrata) — and, primarily, the direction of the cross-stratification dip.

This striking similarity in the structure of the Upper Turonian flysch leads us to believe that flysch teleconnection, i. e., correlation of flysch rhythmograms over hundreds of kilometers, is feasible — provided certain conditions are fulfilled.

The first of these conditions is a detailed study of many, rather than a single, Upper Turonian sections of the eastern Georgia flysch, to ascertain that the sections are correlative over short distances. With a number of such sections at hand, if possible representing the entire zone (from the Kakhetia Range to Southern Osetia), both along and across the strike, one may proceed to correlate the graphs. Facies differences should be taken into consideration, i. e., flysch correlated with flysch, sub-flysch with sub-flysch, etc. This second condition should be strictly adhered to because earlier attempts at flysch connection have shown that correlation is often readily obtained along the strike of facies zones and is quite difficult across the trough [2, 3]. That probably is true also when distances between the sections are considerable.

Only those sections should be correlated whose stratigraphic position is definitely established with relation to the Ananur marker horizon — so that the results can be checked by other methods.

The connection should be supplemented by a statistical analysis of thicknesses for more than a merely visual comparison of sections. An adequate number of rhythms should be studied in detail (at least 200-300). Only then can one be sure that the similarity of the curves is not accidental.

With these conditions fulfilled, and with data from an adequate number of sections, one may proceed with teleconnection. The task is arduous but the results will repay the time and effort because a horizon-by-horizon correlation over distances of more than 1000 km may be of tremendous theoretical and practical importance.

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Received, 28 December 1959

THE PRINCIPAL CAMBRIAN-ORDOVICIAN DISCONTINUITY IN THE NORTH PART OF THE SOVIET BALTIC REGION[1]

by

T. N. Davydova

INTRODUCTION

The position and extent of the discontinuity between the blue shales and the glauconite beds in the northern part of the Soviet Baltic region is subject to argument.

Most geologists, including the author, adhere to the concept of F. B. Schmidt [12, 13] and of those who followed him [6, 15-17, 19, 24]. They all postulate the presence of a long continental hiatus prior to the deposition of sediments containing an Obolus fauna (Pakerort beds) which are regarded as the beginning of a transgression; the underlying beds (the fucoid sandstones of F. B. Schmidt) are supposed to be Lower Cambrian. L. B. Rukhin and B. S. Sokolov present a substantially different interpretation.

L. B. Rukhin [9] regards the lower part of the Obolus sandstones containing the Obolus, and the underlying beds corresponding to the fucoid beds, as a unit (its Sablino formation) in which the Obolus fauna gradually appear; he believes that they were deposited by a regressing sea. He assigns the fucoid beds and some of the Obolus-bearing Pakerort beds to the Middle Cambrian.

B. S. Sokolov [10] presents the following arguments in favor of drawing the Cambro-Ordovician boundary at the base of the glauconite beds instead of at the base of the Pakerort beds:

1) The Obolus beds, in lithology and formation conditions, are more similar to Cambrian sandstones (Izhora) than to the glauconite bed; this is corroborated also by the discovery of Obolidae in the Izhora and Obolus sandstones;

2) "The formation of Dictyonema shales corresponds to the stage of a reduced Upper Cambrian sedimentation" (p. 25);

3) "The 'glauconite sea' transgression is the first major event of regional significance" (p. 25).

Lithological studies carried out by members of the All-Union Institute of Raw Mineral Materials, in 1945-1951, under the direction of Ts. L. Goldstein and the author, provide evidence for solving these controversial problems.

The data cited below are based on a study of a portion of the section between the blue shales and the glauconitic limestones, with the Pakerort beds studied in particular detail (between Pakri Point and the village of Kopor'ye); individual sections were studied along the Izhora, Tosna, and Ladoga Rivers.

Thus these data deal with the Glint belt, from Point Pakri to the Syas River.

Materials by Ts. L. Gol'dshteyn, B. Ye. Antypko, and I. A. Panov have been used in this article, in addition to the author's.

The stratigraphic subdivisions are designated by the names of their type sections used in the correlation. Thus, the west Estonian section was taken as representative of the lower section of the Lower Cambrian in this belt. The names, "lower and upper zones of Eophyton beds" and "fucoid beds", are used in the meaning attributed to them by F. B. Schmidt [22, 23] and A. Öpik [17, 19] for the type section (Kakumyagi Peninsula, Tiskre, Tallin area). A study of specific individual sections has shown that the term "Izhora beds", now used instead of "fucoid beds", has lost its original meaning;[2] accordingly it is mentioned

[1] O meste osnovnogo pereryva v razreze kembriya i ordovika severnoy chasti Sovetskoy pribaltiki, (pp. 58-70).

[2] For instance, B.S. Sokolov [10, p. 22] believes that the Izhora beds are correlative with the Sablino and Ladoga formations of L. B. Rukhin. The same interval is assigned to the Izhora beds in the 1956 Stratigraphic Glossary.

here only with reference to other works which use the two names as synonyms. The term, "Pakerort beds" (horizon) is more or less in good standing. "Glauconite beds" are those occurring between the Dictyonema shales and the Orthoceratites limestones.

For our purposes we discuss in this article the structure of the Pakerort beds, the lower boundary of the Obolus fauna, the nature of the deposition of the Pakerort beds, the significance of their lower boundary with reference to that of the glauconite beds, and evidence for a Lower Cambrian age for the rocks underlying the Pakerort beds.

THE PAKERORT BEDS AND THEIR BOUNDARIES

The structure of these beds is quite complex [5]. Three cycles are identified having a 4-7 meter thickness, each showing a sharp, usually uneven lower contact with the comparatively coarse-grained material. The two lower cycles are represented mainly by fine-grained sandstones and sands with fragments and whole shells of non-articulated brachiopods, with subordinate dark, gray to brown shales.[3] The lower of the two cycles is correlative with the lower Obolus sandstone or the Acrotreta zone of A. Öpik [20, 21] which contains Lingulella, Acrotreta, and Obolus. Deposits of this cycle are restricted to the area of the villages of Iru, Tsitre, Maardu, in western Estonia, and to the Volkhov and Syas Rivers. Over a broad expanse, the base of Pakerort beds usually consists of the middle cycle sandstones (Figure 1); and of the upper cycle in some areas (Toyla, Varva, Summa River). The upper cycle starts with fine- to medium-grained sandstones and sands containing abundant obolid fragments. The sand grains are usually coarser than in the underlying horizons; they gradually change to argillites (Dictyonema shales). Rocks of the middle and upper cycles contain Obolus and Dictyonema. On the whole, they correspond to the Obolus-Dictyonema zone [21].

A. Öpik points out the substantial difference in faunas of this and the Acrotreta zones and assumes that the latter (the lower cycle) is possibly the topmost Cambrian. He assigns to the Ordovician the overlying, widely distributed Obolus-Dictyonema beds (middle and upper cycles).[4] At the same time, considering the presence of forms common to both zones, he regards the Acrotreta zone as the lower member of the Obolus-Dictyonema section deposited during a transgression. Lithologically, rocks of the low and middle cycles are similar.

In the Udria area, the middle and upper cycles constitute a single unit with a gradual transition between them, i. e., without the sharp and uneven intra-Pakerort contact present in most sections. The uneven, distinct contact between the Pakerort beds and the underlying Fucoid occurs everywhere.

Considerable bearing on our problem is held by the obolid fauna distribution, both in the Obolus interval and in the fucoid (Izhora) beds, and the gradual nature of the transition between the two in Leningradskaya Oblast'. L. B. Rukhin's monograph [9] alone presents specific material on that subject (columnar sections and a description of exposures with the formations identified).

Rukhin correlates the Sablino formation[5] identified by him in Leningradskaya Oblast' with the fucoid beds of Estonia and notes that Obolus appears at the top of that formation.

There are no objections to this correlation for most of the sections he describes. There is no mention of Obolus remains in the description of the Sablino formation of these sections. In Estonia they do not occur. There are, however, serious objections to assigning to that formation, rocks described as containing an Obolus fauna. We turn now to sections we have studied along the Lamoshka and Izhora Rivers (at the village of Staraya Myza) and which are located nearer to the Estonian sections (Figure 1).

In these sections, Rukhin assigns to the Sablino formation the light-grey, very fine-grained sandstones with a cross- to wavy stratification and containing fragments and whole phosphatic obolid shells, also subordinate, dark-grey to brown argillites. In the Lamoshka River area, these sandstones belong to the Sablino formation. Here, they rest on an uneven surface of green-grey glauconite siltstones definitely correlated by L. B. Rukhin and the author, with rocks of the Estonian lower Eophyton beds. Assigned to the Sablino formation in the Izhora section is the bulk of the obolid-bearing sandstones[6] which

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[3] Weathering to clays, with a change in color.

[4] In Figure 1, deposits of the middle cycle, overlying the fucoid beds, are assigned to the Ordovician (O_1).

[5] Between the lower zone of the Eophyton beds and the glauconitic beds, L. B. Rukhin identifies (reading up) the Sablino, Ladoga, and Tosna formations.

[6] The upper part of these sandstones (from several cm to one m) has been designated as the Ladoga formation by L. B. Rukhin.

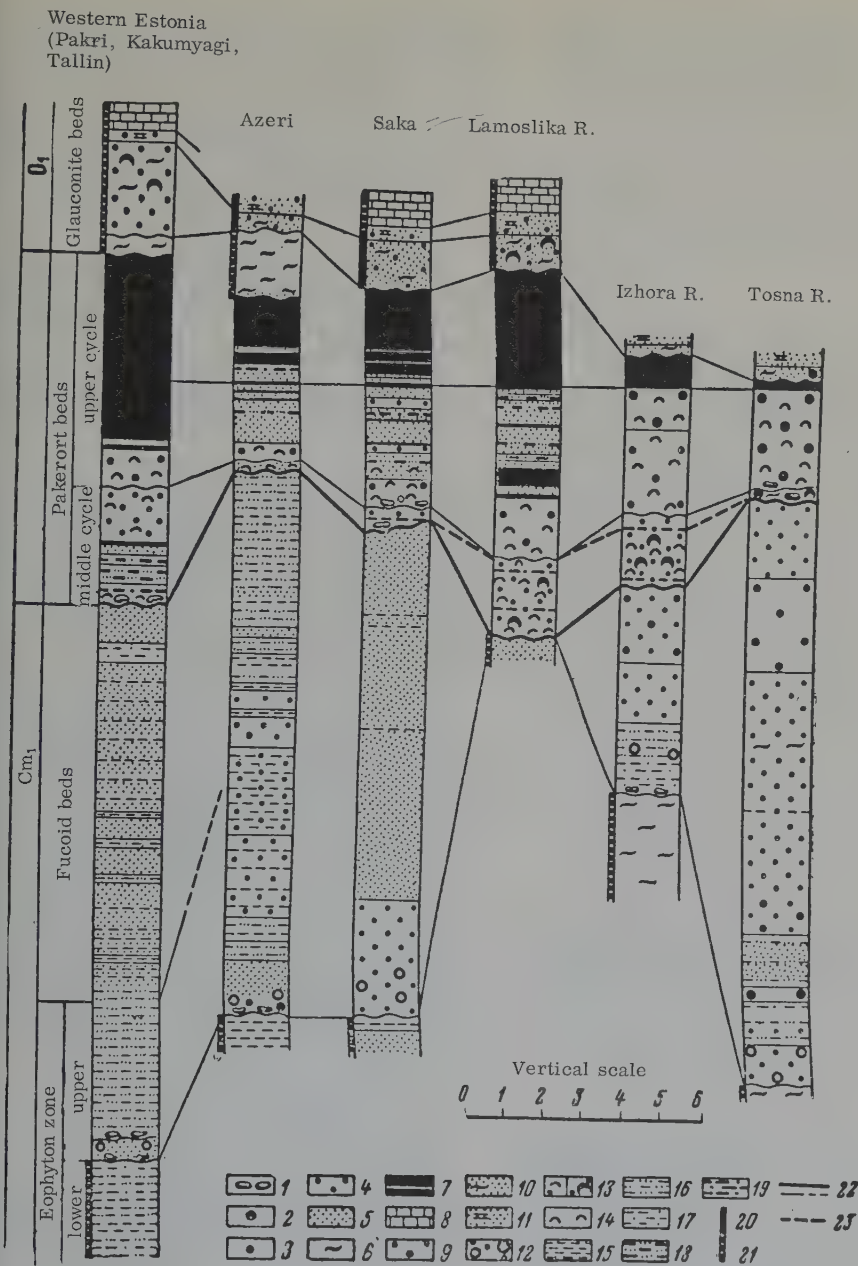

FIGURE 1. Correlation of upper Lower Cambrian and Lower Ordovician sections from Pakri Point to the Tosna River (datum - base of upper horizon of the upper cycle)

1 - gravel; 2-4 - sandstones and sands: 2 - coarse- and medium-grained; 3 - fine-grained; 4 - very fine-grained; 5 - coarse siltstones and silts; 6 - shales; 7 - argillites; 8 - limestones; 9 - about the same proportions of sand grains of similar size; 10 - addition of argillaceous material; 11 - calcareous rocks; 12 - rounded nodules a dolomitic cement; 13 - fragments and whole shells of obolids in rock; 14 - shell detritus; 15 - fine- to medium-grained argillaceous siltstones, silty shales, and shales; 16-19 - alternation of rocks (not to scale): 16 - coarse siltstones (or very fine-grained sandstones) predominant over No. 15; 17 - No. 15 predominant over coarse limestones (or very fine-grained sandstones); 18 - same as 16 but with argillites instead of No. 15; 19 - same as 17 but with argillites predominant; 20 - glauconitic rocks (more than 50% glauconite); 21 - consistent and considerable addition of glauconite (less than 50%); 22 - stratigraphic boundaries; 23 - upper boundary of L.B. Rukhin's Sablino formation.

rest on an uneven surface of a greyish-white, very fine- to fine-grained friable sandstones barren of obolid fragments. These barren sandstones, too, have been assigned to the Sablino formation by L. B. Rukhin.

The upper boundary of the very fine-grained obolid sandstones is the base of the Tosna sandstones of L. B. Rukhin (lower part of the upper cycle of Pakerort beds), which is sharp and uneven.

A progressive correlation of many sections in Estonia and in the western part of Leningradskaya Oblast' shows that these obolid sandstones, assigned to the Sablino formation by L. B. Rukhin, correspond to sandstones with dark argillites and an obolid fauna, from the middle Pakerort cycle of Estonia (Figure 1), usually resting on the fucoid beds. Both of these sandstones occupy the same stratigraphic position, are similar in their rocks, structure of section, the presence of obolids, and differ substantially from the underlying fucoid beds, separated from them by an uneven contact. This correlation has been corroborated by S. N. Naumova's spore analysis which reveals a similarity in spore assemblages from beds we believe to be correlative, and a great difference from those from the underlying beds.

In the eastern part of this region (for instance, along the Lava, Volkov, and Syas Rivers), sandstones with intercalations of grey-brown shales, quite similar to those described above from the middle Pakerort cycle (in composition, boundaries, the presence of whole Obolus shells), and of the same stratigraphic positions, are designated as the Ladoga formation by L. B. Rukhin. Consequently this formation, too, corresponds on the whole to the middle Pakerort cycle.[7] The possibility of the Ladoga sandstones being correlative with lower Pakerort beds was voiced by K. K. Myuyurisepp [7].

L. B. Rukhin does not regard the presence of the Ladoga formation in Estonia as definitely demonstrated [9, p. 163]; however his stratigraphic column [9, p. 168] shows it and the Sablino formation to be correlative with the fucoid beds. The same correlation is reflected in subsequent works (B. S. Sokolov, [10]; Stratigraphic Glossary of the U. S. S. R.).

Another reason for assuming the development of Obolus in the fucoid sandstones was the Popovka River section mentioned in M. E. Yanishevskiy's work [14]. In that section, Obolus remains occur throughout the sands resting on an uneven surface of bluegreen shales. M. E. Yanishevskiy believed that the fucoid sandstones were missing in that area and that the Ungula (Obolus) sandstones (of the middle Pakerort cycle) rested directly on Lower Cambrian shales. This belief has been corroborated by our studies in the area west of the Popovka River.

Thus, there is no evidence to support the appearance of obolids in fucoid beds. All data shows that this fauna appears first in the Pakerort beds.

We shall pause briefly for a description of the lower boundary of these beds.

The lower boundary of the lower Pakerort cycle (the Acrotreta zone, [21]; Yulgaz member, [7]), has been observed only in the Yulgaz area, where it is sharp and fairly even according to K. K. Myuyurisepp. Present at the base of these sandstones are Obolus fragments, also pebbles of the underlying fucoid beds, some of them weathered, as noted by K. K. Myuyurisepp.

Over a long distance from Pakri Point to the Tosna River, the lower boundary of the Pakerort beds is, as a rule, the base of the middle cycle, less commonly of the upper, resting on fucoid beds or on older horizons. In either instance, this boundary is sharp and uneven, while the underlying rocks often contain pockets and joints filled with sand and obolid fragments; there are local, uneven surfaces reminiscent of weathering. All this indicates that this surface, prior to the deposition of Pakerort beds, was formed under continental conditions. More often, the rough spots and the protruding edges of underlying rocks show evidence of wave action. The nature of this surface, and the pebbles of the underlying rocks, suggest that the rocks thus disintegrated were well consolidated and cemented.

A detailed description of this lower boundary of Pakerort beds in various areas of the Baltic region along, with actual field data, is given in a special paper by K. K. Myuyurisepp [7]. His data, as well as ours, show that the concept of a gradual transition from the Pakerort beds to the underlying beds in Leningradskaya Oblast' has no basis in fact. A reason for such misconception, aside from the above-named instances (Sections along the Lamoshka, Popovka, Izhora Rivers), may have been the fact that, locally, the lower Pakerort contact is not sharp, at first glance. Cases have been observed in one exposure between sections where the contact is quite conspicuous.

For instance, a special study of the lower Pakerort contact on a cliff on the right bank of

[7] With the exception of sandstones in the lower part of this formation, along the Volkov and Syas Rivers, probably corresponding to the lower Pakerort cycle.

FIGURE 2. Contact of the Pakerort and fucoid beds
Diagrammatic sketch of an exposure in the right bank of the Izhora R.

1 - white very fine- to fine-grained quartz sandstones, barren; 2 - white very fine-grained sandstones with fragments and whole obolid shells; 3 - layer of brown clay; 4 - boundary between Pakerort and fucoid beds.

he Izhora at the village of Staraya Myza Figure 2), shows that it is definite and sharp here an intercalation of dark-brown clay or mall rounded concretions of brown ironstone est on an uneven surface of white fine-grained ıcoid sandstones. However, where the con-retions are missing and the clay layer wedges ut, these sandstones are overlain directly y very fine-grained greyish-white sand-tones containing small obolid fragments; the ontact between the two sandstones is incon-picuous, although traceable by differences in ieir mechanical composition, color, and aterial composition, as far as the next sec-on where it is again well-defined. Above this ontact there are phosphate fragments of bolid shells; they have not been observed elow it despite a most painstaking search. ne of L. B. Rukhin's columns [9, p. 119] hows a section where this boundary is vague; or this reason it is not shown and the lower art of the Obolus sandstones is assigned to ne Sablino formation.

Even such a classic and unquestionable in-tance of a sharp contact at the base of akerort beds as the Point Pakri exposure ontains sections where the white siltstones re overlain by light-grey very fine-grained andstones containing rare and small obolid emains and where the boundary between them s inconspicuous.

The idea of the "Dictyonema beds having een formed at a low point of Upper Cambrian edimentation" is in contradiction to the vidence. The distribution of all three akerort cycles, as well as the detailed dis-ribution map of the first cycle with its ictyonema shales, and the occurrence of nderlying rocks — all show that the Pakerort ediments were deposited in an expanding asin.

Indeed, out of the three, the lower cycle as a restricted distribution, the middle cycle is wider, and the upper cycle, containing the Dictyonema shales, wider yet. A comparison of the facies distribution maps of the latter [5], more specifically of their position with reference to littoral sediments, shows that the basin expanded gradually so that a given zone received deposits ever further removed from the shore and coming from a progressively leveled land. The maximum transgression, for that particular period, has been established as the deposition time of the upper Dictyonema shales.

A survey of data bearing on Pakerort beds and their stratigraphic equivalents superimposed on various underlying horizons in the Baltic region, Sweden and Norway is given in the works of V. V. Lamanskiy [6] and A. Tammekann [24]. Our own studies have shown that Pakerort beds in the western part of the Leningradskaya Oblast' pass from fucoid beds to the upper- and then to the lower Eophyton zones.

The following data also should be considered. All investigators, including L. B. Rukhin and B. S. Sokolov, correlate the Dictyonema shales of Scandinavia with those of the Baltic region. Consequently, as compared with the underlying deposits whose thickness in those regions is known, the Dictyonema shales are the most consistent and most common, which is in good agreement with the transgressive nature of the Pakerort beds in the Baltic region.

We turn now to the comparison of the lower contact of the glauconite and Pakerort beds.

The lower member of the glauconite beds is represented by pale-green shales up to 2 m thick, containing intercalations of green authigenous glauconite grains (Varangu member [1]). They are preserved in the areas of Azeri, Varangu, Tallin, and Leetse. Their lower contact is sharp and uneven.

Over long distances, the Dictyonema beds are overlain directly by glauconite sandstones and siltstones (Horizon B_1). Their lower contact is sharp, unlike that of the Pakerort beds, it is somewhat uneven, containing mostly flat knobs, up to a few centimeters high, as seen in exposures and clearings. From Pakri Point to the Narva River, the glauconite sandstones pass from the upper part of the Dictyonema shales to their lower part, i. e., truncating an interval up to 6 m thick, in 250 km.

A characteristic feature of the lower boundary of glauconitic sandstones, setting it apart from that of the Pakerort beds, is the numerous sinuous trails of mud-eaters,[8]

[8] Mud-eaters' trails are present in the lower part of Pakerort beds but do not continue below their base.

starting at the contact and piercing the upper 10-15 cm of the Dictyonema shales. The hollow tubes are filled with glauconite sands or with silts from the overlying layer. The extensive occurrence of these trails, and as the nature of the argillite inclusions in the glauconite sandstones, suggested their submarine origin at a period when the future Dictyonema shales were a habitat for mud-eaters.

This assumption is in agreement with the general distribution of glauconite Pakerort facies. For instance, coarser-grained terrigenous rocks are developed in the glauconite beds wherever coarse-grained, littoral Pakerort deposits are present.

Thus, there is a reason to regard the lower glauconite bed contact in this area as the result of shallower depths and of submarine erosion, rather than of emergence. It should be kept in mind that erosion was in ascendance over deposition after the Dictyonema shale deposition, as well as after the deposition of glauconite shales which are the low member of the glauconite beds.

To evaluate the importance of these boundaries it is necessary to determine the duration of their corresponding sedimentary hiatuses (either submarine or continental). As demonstrated above, the very nature of the lower contacts of the Pakerort and glauconite beds, and of rocks at their base, points to their substantial difference in that respect. The Pakerort beds were deposited on rocks already consolidated, while the glauconite beds were deposited on viscous ooze rich in organic matter — subsequently lithified to the argillites of the upper Pakerort beds. This suggests that the period between the deposition of the fucoid and the Pakerort beds was longer than that between the deposition of the Pakerort and the glauconite beds.

This conclusion is in accord with S. N. Naumova's data to the effect that the composition and size of spore assemblages change radically above the lower contact of Pakerort beds, while these changes are insignificant in the transition from the Dictyonema shales to the glauconite beds.

In advocating the transfer of the Cambro-Ordovician to the base of glauconite beds, B. S. Sokolov states, "The most potent argument for this new boundary is the fact that the Ordovician transgression in the Soviet Baltic region is not represented by its oldest stages: the glauconite sandstones are not as old as those Ordovician rocks known in Scandinavia. Missing in our section are the top of the Dictyonema and the Ceratopyge beds (top of zone 2-e and all of zone 3-a, corresponding to the known Tremadoc of England)" [10, p. 24]. He believes this to be due to the "total absence of sedimentation at the onset of the Ordovician." The data cited above are in contradiction with this conclusion.

The presence of the Tremadoc in our section is open to argument. On the basis of her analysis of data from English, Norwegian, and Swedish sections, T. N. Alikhova [3] assigns the Pakerort beds to the lower Tremadoc and the glauconite beds to the upper.

It should also be kept in mind that the oldest glauconite beds in our section are shales underlying the glauconite sandstones. These shales have not been studied paleontologically. Also inadequately known is the fauna from the upper Dictyonema shales of Western Estonia, where their section is more complete. A more complete sequence of faunal zones may be established as the result of a special study. This will determine more accurately the standard section interval corresponding to the boundary between the Dictyonema and glauconite shales of the Baltic section.

It should be noted that Ye. A. Reytlingen identified in 1946, from her study of thin sections, foraminifera in the upper Dictyonema beds (the upper member, better developed in western Estonia) similar to representatives of the genera Psammosphaera and Thurammina, described from the Silurian of America by Ireland in 1939, and by Dunn in 1942.

THE AGE OF ROCKS UNDERLYING THE PAKERORT BEDS

The conclusions concerning a long continental sedimentary hiatus prior to the deposition of the marine Pakerort beds is supported by data indicating a Lower Cambrian age for the underlying fucoid rocks.

There are no published objections to the Lower Cambrian age of the upper Eophyton beds underlying the fucoid beds. That age has been established by A. Öpik [17, 19, 21] by comparing the sequence of rocks and faunas in Western Estonia, Sweden, and Norway; more specifically, he mentioned the presence of Mickwitzia monilifera in both the upper and lower zones of the Eophyton beds.

Our own study has shown that rocks of the upper Eophyton zone and of the overlying fucoid beds in the belt from Point Pakri to the Tosna River, represent a single complex resting with an uneven contact on the lower zone (Figure 1). This idea is not new. Even A. Öpik stressed the arbitrary nature of the boundary betwen the upper Eophyton and fucoid beds of Western Estonia and the impossibility of their "differentiation by petrographic criteria" [18].

.. B. Rukhin, in his correlation of sections
n Leningrad Oblast' and Estonia, proposed
ombine the fucoid sandstones with Scenella-
ring shales [9, p. 165].

. Öpik draws the lower boundary of the
er zone on the base of very fine-grained
y sandstones, up to 2 m thick, which con-
small rounded concretions cemented
olomite ("dolomitic sandstones with
erical concretions" of Mickwitz). These
lstones rest with an obviously uneven con-
on argillaceous rocks of the lower zones
aining authigenous glauconite. Upon
thering, the concretions develop a film of
vn oxides and stand out in small spheres
solitic sandstones"). Upward, they change
lually to light-grey to white, very fine-
ned sandstones and siltstones with thin
ged-out intercalations of silty shales.
ording to A. Öpik [17], the lower part of
interval contains a fauna of Mickwitzia
ilifera, Scenella discinoides, and
kwitzia formosa.

n accordance with the disappearance of this
a, he determined the upper zone to be
m thick. Occurring in sandstones contain-
dolomitic concretions in its lower part,
spotty and thin (0.02-0.1 m) conglomerates
mall argillaceous sandstone pebbles con-
ing Mickwitzia monilifera. They are
sing in many sections and the base of the
er zone is represented by sandstones con-
ing dolomitic concretions.

Described below is a section of the upper
e on the Kakumyagi Peninsula (after A.
k, with some details of our own). Resting
ypical, lower zone rocks, it is as follows
ding up):

) Conglomerate of small argillaceous
lstone pebbles coated with a dark crust;
kness, 0.02-0.1 m.

2) Very fine-grained sandstones or coarse
stones containing small round concretions
to 1 cm in diameter), cemented with
omite and coated with limonite; 0.5-2.0 m.

3) Conglomerate, similar to that of member
, 0.02-0.1 m. Members 1-3 contain
kwitzia monilifera.

4) White to light-grey, very fine-grained
dstones and subordinate silty shales with
nella discinoides and Mickwitzia monilifera;
m.

These change gradually to white and light-
y siltstones containing shale intercalations
oid beds).

The above-named fauna of the upper (Scenel-
zone has not been observed east of Kunda.
A. Öpik [10] explains this by a westerly retreat of the sea. He adds, however, that he leaves open the question of a wedging-out, east of Kunda, of rocks correlative with the fossiliferous rocks of western Estonia. We now turn to our own data bearing on this subject.

In the Glint escarpment, at the village of Merekul, in the Azeri area, the following section rests on the lower zone with a sharp contact:

1. Conglomerate of small pebbles of a grey argillaceous silt, coated with a dark crust and cemented with light-grey, very fine-grained sandstone; 3-4 cm thick.

2. Very fine- to medium-grained sandstone, silty, with typical small spherical concretions (up to 1 cm) of dolomitic sandstone. These concretions often merge into bizarre reniform bodies. The weathered surface is warty ("pisolitic sandstone"); 0.2-1.0 m.

3. Coarse white siltstones and very fine-grained sandstones containing numerous wavy contacts covered with small flat pebbles of blue-green silts. Present in the lower part are scattered flat pebbles similar to those of member one; 1.25 m.

Occurring higher up, with a gradual transition, are very fine-grained white sandstones changing to coarse siltstones with argillaceous intercalations; thickness, about 12-13 m.

No fauna has been observed.

The sequence of members 1-3 is similar to that from the lower part of upper Eophyton beds on the Kakumyaga Peninsula, described above.

The conglomerate beds (member one and the base of member three) are traceable for only short distances. Sandstone containing dolomitic concretions ("pisolitic"), usually limonitic on the surface, is present everywhere in the Azeri and other sections of eastern and western Estonia.

Thus, developed in the lower layer of the rocks resting with an uneven contact on the lower Eophyton zone east of the Kunda River, are rocks which contain, west of the Kunda, a marine Lower Cambrain Scenella fauna.

East of Azeri, we traced these same rocks and their gradual transition to the overlying beds, correlative with the fucoid beds of Western Estonia, in a number of sections (areas of the villages of Sak, Sillamyae, the city of Narva, Luga River).

It should be noted that A. Öpik [19] described Diplocraterion remains from the sandstones

overlying the lower Eophyton zone and corresponding to the upper zone in a section at Narva on the river. Similar remains were noted by K. K. Myuyrisepp [7] in a section along the Narva and the Luga, near Kingisepp in the upper section of the overlying rocks corresponding to the fucoid beds.

Only the lower part (about 1.5 m) of sandstones containing occasional limonite concretions (upper Eophyton beds) is present in the Suma River sections east of Kingisepp. Near the station of Kotly, Diplocraterion was identified by K. K. Myuyurisepp [7], from the sandstones in the upper Eophyton beds (0.1 m below the base of the Pakerort beds).

Farther east in the area of the Lamoshka River, the village of Kopor'ye, and the Popovka River, the Obolus sandstones of the Pakerort beds, occurring above a sharp erosional discontinuity, change directly to siltstones and shales of the lower zone (the Vollortella tenuis zone of the Eophyton beds).[9]

Still farther east, in sections along the Izhora and Tosna Rivers, rocks characteristic of the lower part of the western Estonian Scenella zone reappear in the corresponding interval of the stratigraphic section (Figure 1). As in western Estonia, they are connected by quite gradual transitions with the overlying rocks corresponding to the fucoid beds. They are particularly well represented in a section on the left bank of the Tosna (in the Pustynka area), by coarse friable siltstones containing rounded dolomitic concretions. These siltstones rest with a sharp contact of the blue-green shales of the lower Eophyton beds.

According to Ts. L. Gol'dshteyn, similar rocks occur in the corresponding interval of a section on the Ladoga River.

Thus, rocks typical of the lower part of the Lower Cambrian Scenella zone, most persistent in western Estonia, are also traceable over long distances east of the village of Kunda as far as the Ladoga River. They are connected everywhere by gradual transitions with rocks correlative with the fucoid beds of western Estonia and are separated by a clean-cut boundary from the lower Eophyton beds. No erosional break comparable to that at the base of the upper zone has been observed in the overlying rocks, as far up as the base of Pakerort beds, where only local discontinuities are present between the cross-bedded sandstones and argillaceous rocks. Similar contacts are also present within the lower Eophyton zone.

Authigenous glauconite, a mineral typical of the underlying lower zone rocks, does not occur in the sections comprising the upper Eophyton and glauconite beds.[10]

Thus, rocks of the upper Eophyton zone and the fucoid beds which are similar to them and connected to them by a transition, indeed represent a single complex. Its lower part, in western Estonia, contains a Lower Cambrian index fauna of the Scenella zone; east of there, it contains remains of Diplocraterion which also occur in its upper part.

A. Öpik [19, 21] has presented evidence that the Scenella zone and the overlying fucoid beds are correlative with the Holmia kjerulfi zone of Sweden and Norway, i. e., not the uppermost Lower Cambrian zone of those regions. All this confirms a Lower Cambrian age of the fucoid beds — a conclusion arrived at by F. B. Schmidt [22, 23], and by A. Öpik [17-19], M. E. Yanishevskiy [14, 15], and other investigators.

B. S. Sokolov presents the following arguments for a Middle Cambrian age of the fucoid (Izhora) beds: "The presence of a clean-cut discontinuity between the Lower and Middle Cambrian, in Sweden (a province nearest the Soviet Baltic region) suggests that the discontinuity between the Izhora and Eophyton beds, and between the Izhora and "blue shale", likewise marks the Lower-Middle Cambrian boundary" [10, p. 23].

As shown above, there is no discontinuity between the Izhora and the underlying beds. The Izhora (fucoid) beds are gradually replaced by upper Eophyton beds, while a clean-cut, uneven contact is traceable lower in the section at the base of the upper Eophyton zone. A Lower Cambrian age of the latter is agreed upon by everybody, including B. S. Sokolov, so that there is no reason to assign a Middle Cambrian age to the fucoid beds.

As to the upper boundary of the upper Eophyton beds, it cannot be regarded as the expression of any substantial hiatus. Indeed, these beds, where preserved, are underlain by the lower zone and in western Estonia, contain a common Mickwitzia monilifera fauna.

This boundary, like the erosional surfaces with the mud cracks in the lower Eophyton beds, is a feature of the Lower Cambrian section in the northern part of the Soviet Baltic — shorter than Swedish sections, as noted by A. Öpik [19]. Another regular feature is that the

[9] Our studies have shown that the alternating sandstones, siltstones, and clays of the lower zone east of the village of Kopor'ye change to shales, thus corroborating one of M.E. Yanishevskiy'a assumptions [15].

[10] The lower parts of the upper zone locally contain glauconite but are obviously redeposited.

us, initiated after the deposition of our ver Cambrian section, was comparatively rt in Sweden, while lasting at least through entire Middle Cambrian in the extreme thern part of the Baltic province of less nplete sections — until the onset of a sea nsgression containing the first obolid fauna. re is a general agreement that the age of se deposits is not older than the top of the per Cambrian.

The concept of an intensive erosion of rocks erlying the fucoid beds — a hiatus especially spicuous in Leningradskaya Oblast' [4, — has originated for a number of reasons. r instance, the Obolus sandstones of the erort beds, indeed resting erosionally on erent underlying horizons, were arroneous- ssigned to the fucoid beds in some sections Lamoshka, [9]; R. Popovka, [8]). Every- disregarded the fact that L. B. Rukhin had related his Sablino formation not only with oid beds of Estonia but also with the nella zone [9, pp. 165, 168). Also disre- ded was the gradual transition of alternating illaceous, arenaceous, and silty lower hyton beds of Estonia to shales, as occurs western Leningradskaya Oblast'. Because his, the fact of these shales being overlain Lower Cambrian rocks, observed east of or'ye, was regarded as an equivalent of the er rocks overlying the "blue shales" B of F. B. Schmidt).

The data cited show that the upper Eophyton e should be combined in a unit with the oid beds rather than with the lower Eophyton e, as is done in the present stratigraphic umns (the 1958 Estonian and that of B. S. olov, [10, 11]).

CONCLUSIONS

There are no data refuting the basic concepts the history of the development of the north- part of the Soviet Baltic in the Cambrian early Ordovician as established by F. B. midt, V. V. Lamanskiy, A. Öpik, and M. E. ishevskiy. On the contrary, all available terial indicates those concepts to be basical- correct.

A continental hiatus preceded the deposition he Pakerort beds and embraced the top of wer Cambrian, all of the Middle Cambrian, l most if not all of the Upper Cambrian. The erort beds represent the onset of a trans- ession following this long sedimentary hiatus.

Opinions differ on the more precise dating this transgression, i. e., on the correspond- interval in the standard section of England. s either the top of the Cambrian or the base the Ordovician. Additional study is neces- y for a more definite answer. In any event, the arguments presented here for the position of the principal sedimentary hiatus in the Baltic section should be taken into considera- tion.

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Received, 15 July 1960

RELATION OF THE PRESENCE OF COAL TO THE FACIES OF PEAT ACCUMULATIONS AND THE ENCLOSING SEDIMENTS IN THE ORSK COAL BASIN[1]

by

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There are many references in the geological literature on the relationship between petrogenetic coal types, the character of the coal measures, and the lithofacies composition of the enclosing rocks [4]. However, the more or less specialized studies in that field have been carried out largely in the last decade. These are dealt with in varying degrees of detail in the materials on the various deposits as treated in the works of Yu. V. Stankevich [8], P. P. Timofeyev [9, 10], L. P. Nefed'yeva [5, 6], M. I. Ritenberg [7], L. I. Bogolyubova, and V. S. Yablokov [2], I. B. Volkova [3] and others.

This article presents some of the results of a comprehensive study of the presence of coal, the petrographic and facies composition of coals, and the facies of the enclosing Middle Jurassic rocks in the Orsk basin. It is not the authors' purpose to go into all of these details but merely to determine the quantitative relationship between the presence of coal (number and thickness of coal seams), the lithofacies composition of the coal measures, and the petrogenetic types of coals. This study was carried out in connection with the 1956-1959 geological exploration in the Orsk basin by the Lower Mesozoic Expedition of the Coal Laboratory of the Academy of Sciences of the U. S. S. R., under the direction of I. I. Gorskiy and N. I. Leonenok.

The lower Mesozoic coal measures of the Orsk basin, on the east slope of the Northern Urals, fill a large depression in the Paleozoic basement. This depression extends meridionally along the left bank of the Ora River, on its middle course. The Orsk basin proper is associated with the central part of the depression; its basic coal mining areas (near the village of Mamyt) are located 65 km south of Orsk and 120 km east of Aktyubinsk. The coal occurs in Middle Jurassic sediments overlying Lower Jurassic areno-argillaceous alluvial to lacustrine deposits and are overlain by Tertiary and Quaternary deposits. The basin is somewhat asymmetric. Its western half is characterized by a more complete section without the post-Jurassic erosion which eroded the upper part of the section in the eastern half. The western boundary of the basin is marked by a major fault; its eastern boundary is erosional, coinciding with outcrops of the lower Middle Jurassic beds. In the south the basin is bounded by a basement uplift which separates its central and southern sections. The northern boundary is drawn arbitrarily on a line where the workable coal measures wedge out. The deposits are virtually horizontal, having only a slight dip toward the center of the depression. Deformations, such as dip steepening, step-like structures, and faults, are only occasionally present along the edge of the depression and near the inner basement uplifts.

The Middle Jurassic coal measures, 250 m thick on the average, consist of alternating, usually poorly cemented very fine-grained shaly to sandy silts, coals, and carbonaceous shales, largely lacustrine to marshy-lacustrine. Several facies groups have been identified, represented by the following principal types:

Type I: lacustrine facies, away from the littoral section. Horizontally finely-stratified claystones containing a fresh-water pelecypod fauna and rare and small plant remains;

Type II: littoral lacustrine facies, also those of small flowing lakes and lacustrine deltas. A rapid alternation of shaly silts and less common fine sandy beds with a wavy to horizontal stratification and containing plant remains in various degrees of preservation;

Type III: Swamp facies. The coal beds are associated with these. Three principal types of fossil peatbogs have been identified:

a) slightly inundated forest swamps. Coals dull to semidull, striated, locally banded, consisting largely of fusainized to slightly fusainized wood tissue; low mineral content;

[1] Svyaz' uglenosnosti s fatsiyami torfonakopleniya i vmeshchayushchikh osadkov v Orskom ugol'nom basseyne, (pp. 71-80).

b) periodically-flowing forest quagmires. Coals dull to semidull, homogeneous to sparsely striated, consisting largely of fine shreds of fusainized tissues with evidence of gelling. The mineral content varies from 10-15 to 30-40%; accordingly, comparatively low-ash earthy and high-ash dense varieties of semifusainized coals are differentiated;

c) stagnant and flowing quagmire marshes. Coals semi-lustrous, lenticularly banded, and striated, containing a large amount of gelled tissues and a low ash content (stagnant bogs). Semidull to dull coals, striated, high-ash, with largely gelled organic remains, and lignites (flowing marshes).

Type IV: facies of stagnant, overgrown basins and marshlands. Silty clays, dark, non-stratified, with root remains;

Type V: alluvial facies of unconsolidated sandstones, fine- to medium-grained, poorly sorted, cross-bedded.

The coal measures in the basin contain a maximum of up to 25 coal seams and interbedded coal, with up to seven workable seams (over 0.7 m thick). The seams vary in thickness, are largely thin (0.1-1.0 m) and average (1-3 m). Thick seams (more than 3 m) are few and areally restricted. Structure of the seams in both simple and complex thin beds are simple, the thicker ones, more complex (two to five coal seams alternating with argillaceous and carbonaceous rocks; the coal seams themselves are not homogeneous, consisting of several coal types). The composite thickness of the coals in the basin is up to 12 m.

As a rule, the coals occur in argillaceous and carbonaceous rocks, with root remains often present directly below the coal seams. Evidence of coal seam erosion is quite rare.

These coal measures are characterized by a definite and regular pattern of interrelated changes in the presence of coal, lithofacies of the enclosing rocks, and the composition of coal throughout the section, and especially in areal extent.

Changes throughout the section are expressed in the principal mining areas of the basin by a gradual upward increase in the amount of coals, carbonaceous shales, and genetically related sediments (Types III and IV), to a maximum approximately in the center, followed by a gradual decrease to zero. The increase in coal is due to an increase in the number and thickness of coal beds. This is accompanied by a change in the coal composition: the proportion of low-ash fusain coals increases in thick seams as compared to the thin ones where semifusainized and clarainized coals predominate. At the same time, the reverse is true for lacustrine deposits of Types I and II. In other words, a maximum content of coal and carbonaceous rocks corresponds to a minimum of lacustrine deposits, and vice versa. This general pattern is modified by local sedimentary features. Thus, the lower part of the section shows a predominance of Type II lacustrine deposits (littoral sediments, those of shallow, flowing lakes, and of lacustrine deltas), while deposits of Type I (farther away from the shore) are in marked ascendance in the upper half. In addition, the lower part contains a small amount of Type V alluvial deposits.

Areal changes are expressed by a definite zonation: 1) in the lithofacies composition of the enclosing sediments; 2) in their coal content; and 3) in the genetic types of coals. These changes can be studied along cross-sections A-B and C-D, through the eastern and western halves of the depression (within the basin). Each section shows four control points (1-8), each representing an average of several neighboring boreholes. The positions of the cross-sections and control points are shown in Figures 1 and 2; the quantitative relations of the various coal and rock facies, in Figure 3. The small number (eight) of the control points is for simplicity of the graph. The description and outline of the zones are based on a great number of field data. Of principal interest are those areas of coal isopachs facing the center of the depression, because it is here that the genetic wedging out of coals takes place (while the section thickness is maintained). Areas of isopachs along the edge of the depression reflect mostly the present coal content formed as a result of a post-Jurassic erosion.

Three principal zones may be designated by lithofacies of the enclosing rocks, the coal content, and types of coals and coal facies — with consecutive single-direction changes from one to another.

1. Zone of littoral lacustrine facies, those of shallow flowing lakes, and lacustrine deltas (Type II — 40-50%), with some alluvial facies (Type V — up to 10-15%), and a considerable proportion of swamp facies, those of stagnant overgrown basins, and marshlands (Types III and IV — 20-45%; Points 1, 2, 5, 6).

This zone is associated with the southern part of the basin, near the basement uplift (more precisely, on its "slope"). As seen from Figures 1-3, it presents an assortment of mostly "continental" sediments in the basin. The proportion of Type I lacustrine deposits does not exceed 20%. The Type I zone is characterized by the highest coal content, containing up to 18-25 coal seams and intercalations having a total thickness up to 10-12 m. There are 5-7 workable seams. The seams are thickest

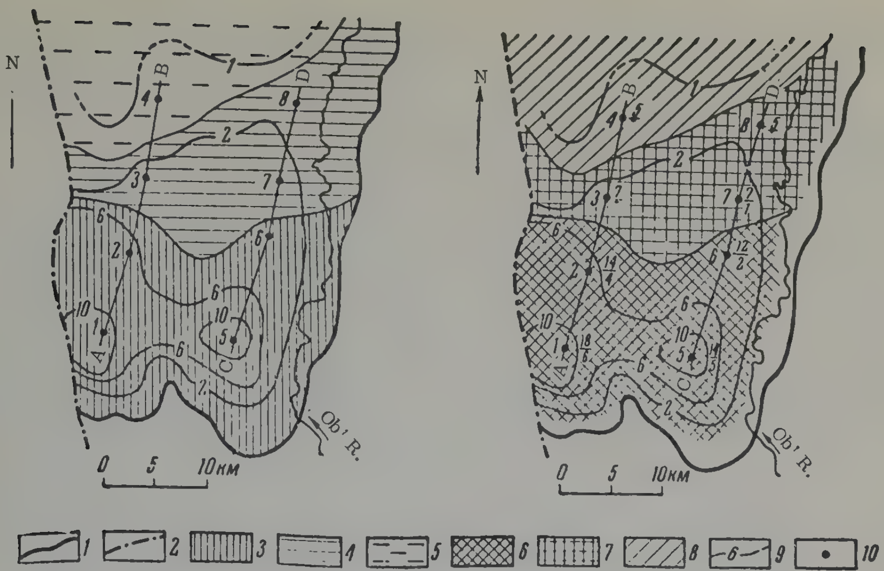

FIGURE 1. (left). Distribution of the predominant facies of the enclosing rocks

FIGURE 2. (right). Distribution of predominant coal facies.

1 - boundary of Middle Jurassic deposits; 2 - normal fault; 3 - zone of Type II lacustrine deposits with some alluvial facies (Type V) and a considerable development of marsh facies, stagnant overgrown basins, and marshlands (Types III and IV); 4 - zone of predominantly lacustrine facies II and I with a slight development of Types III and IV; 5 - zone of predominant Type I lacustrine facies; 6 - zone of slightly inundated forest marsh facies (III-a); 7 - zone of periodically-flowing forest quagmire facies (III-b); 8 - zone of quagmire facies (III-c); 9 - isopachs on total coal thickness; 10 - control cross-section and its number (numerator - the total number of focal seams; denominator - the number of seams over 0.7 m thick).

in the basin (from 2-4 to 6-8 m), the most consistent, and have the lowest ash content. It is here that the principal mining districts are: East-Uralian (Point 1), Pervomaysk (Point 5), and Orsk (Points 2 and 6).

The structure and facies of coal seams in this part of the basin are fairly diversified, with fusains predominant on the whole (70-80%). The latter are 47-48% dull fusains with a fragmentary microstructure and belonging to the facies of forest, slightly-inundated marshes (Type III-a); 23-30% of dull, comparatively low-ash, earthy varieties of semifusains from facies of periodically flowing quagmires (Type III-b); and subordinate (22-30%) gel varieties represented largely by semi-lustrous low-ash types which originated in stagnant quagmires.

A typical feature of the Orsk gelled coal varieties is their high content of plant stalk tissue, with only a small amount of amorphous gel substance. Present as associates are fusainized plant remains whose content often reaches 25-30%. This appears to be due to the fact that the original material accumulated within a poorly differentiated area, under conditions intermediate between aerobic and anaerobic. Because of this, even slight physico-geographic changes resulted in the development of different processes of the transformation of organic matter — fusainization and gelification — with the first one predominating.

Petrographic study has established a definite relationship between the composition and the thickness of coal beds. The workable seams consist mostly of dull to semidull fusains from the facies of slightly inundated forest marshes, with subordinate low-ash gel facies of stagnant inundated quagmires. The predominant component of thin seams is attritus coal having a high mineral content: dense argillaceous fusains of the facies of periodically flowing forest quagmires, and ash-carrying clarain and lignitic coals of the flowing marshes facies. This relationship between the composition and thickness of coal beds has been noted in the literature. I.I. Ammosov [1] has established in the Kuzbas that "a low degree of inundation, the greater thickness of coal seams, and their increasingly allochthonicity, naturally go together" (p. 19).

It appears that this combination of features prevails in the Orsk basin, as well.

The presence in Zone 1 of the thickest coal beds with the highest content of high fusain coals indicates a slow subsidence rate for the deposition area with prevailing oxidation conditions. The latter is corroborated by chemical data, such as the low hydrogen content in the coals (less than 5% of the combustible bulk) and the predominance of alumosilicates ($Al_2O_3 + SiO_3$, up to 80%) in coal ash. The petrographic features of Zone I coals, along with the presence of vertical root remains in the soil and the gradual transition of argillaceous rocks to coal beds, suggest a definitely autochthonous accumulation in forest and quagmire swamps (Type III, a and b).

2. Zone of predominantly lacustrine facies of Types II (40-50%) and I (20-45%), with a small section of a swamp facies; stagnant overgrown basins; and marshland facies (Types III and IV — 10-20%; Points 3, 7, 8).

This zone is located farther away from the southern boundary of the basin, thus differing from Zone I in its considerable development of Type I lacustrine sediments, also in the marked reduction of Types III and IV coals and argillaceous deposits. Only five to ten coal seams and intercalations are present here; they have a total thickness up to 2. 5 m. The coal seams are not over one meter thick, with but a single workable one. They are inconsistent and have a high ash content. This zone includes the southern and eastern sections of the Romankul district (Points 3, 7, 8).

This overall facies change in Zone II has resulted not only in a reduced number and thickness of coal beds but in a change in their petrographic composition. A different ratio of genetic coal types prevails here, with the leading role belonging to the semidull and dull semifusain coals from the facies of periodically flowing forest quagmires (Type III-b), amounting to 44-46% of the total. As shown in Figure 3, the increase in the content of semifusain coals is associated with high-mineral varieties. This may be explained by the fact that plant material accumulated here in areas transitional between forest and quagmire marshes where the changing relief brought about a progressive inundation. It is reasonable to suppose that such areas were subject to periodic floods which brought in additional mineral matter.

Also well developed in Zone II were coals of the stagnant and flowing marsh facies (Type III-c), accounting for 30-38% of the total. The comparatively low-ash fusain coals with a fragmentary structure are subordinate (Type III-a — 18-25%). Thus the vegetation material of Zone II was accumulated in an area more abundant in water, often flowing, so that allochthonous elements appear among the prevailing autochthonous elements.

3. Zone of predominantly lacustrine facies away from the shore (Type I — over 60%; Point 4).

This zone includes the northern part of the basin, a lake during almost all of Middle Jurassic time.[2] The total coal and argillaceous facies of marshes, overgrown basins, and marshland (Types III and IV) do not exceed here 5-10%, and the coals proper are represented by 2-5 thin lenticular, high-ash seams, with a total thickness up to 1. 2 m. There are no workable horizons. The preponderant coal facies are those of the flowing quagmire swamps (Type III-c): semidull high-ash clarains and lignites (57%). The content of low-ash clarain coals formed in stagnant marshes is not over 8%. The fusains are subordinate, being represented largely by high-ash, semifusain, attritus varieties (Type III-b). Typical fragmentary fusain coal facies of slightly inundated marshes (Type III-a) are present in small amounts (15%).

The common features of Zone III coals are their high mineral content (argillaceous material, grains of feldspars and quartz, rock fragments) and an abundance of plant remains, considerably fragmented and gelled. The coarsely fragmented coals (lignites) consist mostly of coniferous xylem, known to be decomposition-resistant.

Thus, the available data indicate an unstable depositional environment for Zone III beds, with a considerable addition of allochthonous elements. The thickness of its coal seams, along with the high content of competent and incompetent gelled elements, suggests a relatively rapid subsidence rate as compared with the southern portions of the basin.

This description of facies zones is greatly generalized and should not imply that these zones are as homogeneous in their rock and coal composition as represented in Figure 1 and 2. It has been mentioned above that only the predominant coal facies are represented. Diagrams in Figure 3 show that almost all types of rocks and coals of the basin are represented in each zone. A more detailed differentiation would be into subzones of predominant facies, of a secondary importance on the whole. For instance, the best known Zone I contains, in its eastern part (Pervomaysk district), a relatively high content of alluvial facies and virtually no Type I lacustrine deposits. In

[2] This lake basin continues farther north as far as the latitude of Orsk, where the content of Type I lacustrine deposits in the Middle Jurassic section attains 80-90%.

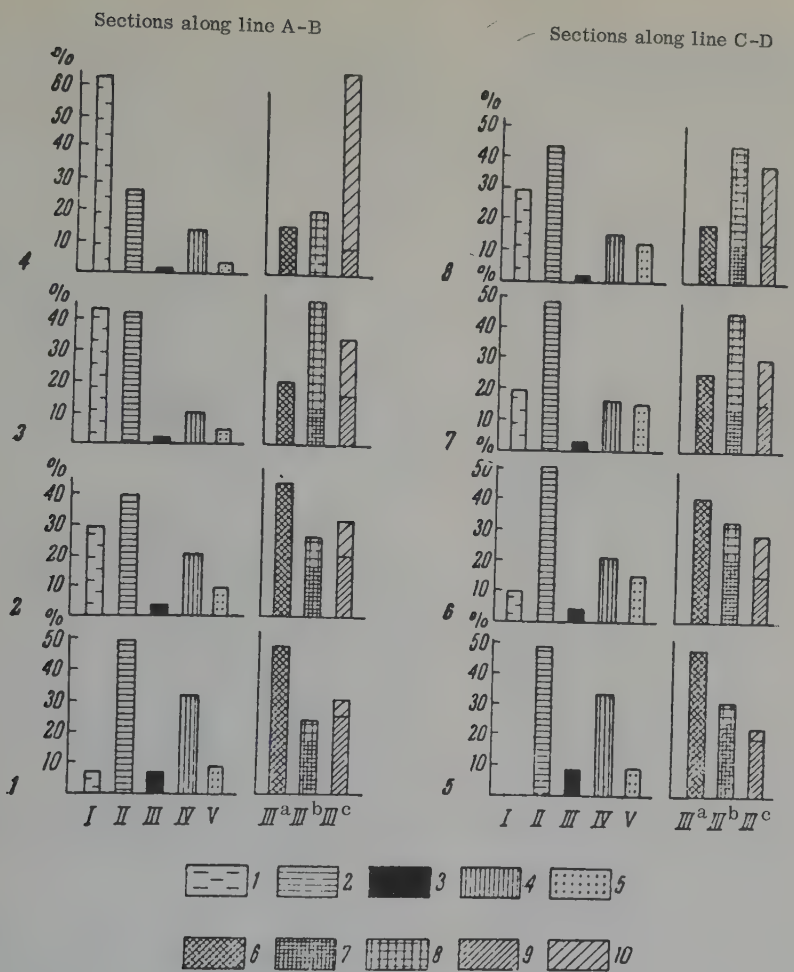

FIGURE 3. Ratios of rock and the coals of various facies

1 - facies away from lake shore. Horizontally-stratified shales containing a fresh-water fauna (Type I); 2 - littoral lacustrine facies, those of shallow, flowing lakes, and lacustrine deltas: rapidly alternating shaly and silty rocks, horizontally-wavy to wavy stratified, containing plant remains (Type II); 3 - marsh facies: coal and carbonaceous shales (Type III); 4 - facies of stagnant overgrown basins and marshlands. Unstratified silty shales containing root remains (Type IV); 5 - alluvial facies. Poorly sorted cross-bedded sandstones (Type V); 6 - facies of slightly inundated forest swamps. Fragmentary fusain coals (III-a); 7, 8 - facies of periodically-flowing forest quagmires. Semifusain attritus coals (III-b), comparatively low-ash (17) to high-ash (18); 9-10 - facies of quagmire swamps, clarain coals (III-c): low-ash, from stagnant marshes (9) and ash-bearing, with evidence of their allochthonous origin in the flowing marshes (10).

ldition, it contains more of the thicker seams ıd correspondingly the highest proportion of sain coals with a relatively low ash content.

The western part of Zone I (East Uralian strict) is poorer in alluvial and higher in ʏpe I lacustrine deposits. The number of icker coal seams is, on the whole, smaller an in the Pervomaysk district, and gelled als play a considerable, although subdinate, role. In addition, there is within ıs zone a narrow belt of thinner coals, predominantly semifusain and clarain high-ash attritus coals and lignites. This belt extends in varying widths along the outer margin of the depression, fringing the main area of the preferential development of low-ash fusain coals. Thus, the complete picture of facies distribution is rather complex, with some of its details unknown as yet. However, the quantitative changes in the principal genetic types of sediments are fairly definite, thus allowing identification of the three zones.

An analysis of data along the two profiles shows that all three elements — facies of the enclosing rocks, the coal content, and the coal facies — change more or less concurrently and are traceable in both the eastern and western halves of the depression. This suggests a regular, rather than a randon, nature of these changes and close relationships between the processes of sedimentation and coal-formation.

The main pattern is expressed in the gradual (upward and laterally) and (from south to north) transition from more "continental" to more "basinward" lacustrine sediments. This is particularly conspicuous in areas where there is a progressive increase in these deposits from a comparatively small section to where they occupy practically the entire section. This is accompanied by a regular reduction in the coal content from a maximum (18-25 seams and beds with a total thickness up to 12 m) in the south, to the insignificant coal shows in high-ash lenses not over 0.5 m thick in the north.

Changes in the enclosing rocks and in the coal content are accompanied by corresponding changes in the petrographic and facies content of coals. This corroborates what Yu. A. Zhemchuzhnikov stated back in 1940 that, despite the differences in the sedimentation conditions of coals and the enclosing rocks, the formation pattern of coal measures provides a framework for the coal distribution pattern [4]. Indeed, the several zones of this area, differentiated by lithology and coal content, contain coals of definite petrogenetic characteristics determined by a number of factors, and whose areal development is related to the overall sedimentary conditions.

For instance the northerly change from a lacustrine marsh to a typically lacustrine peat-forming environment is reflected in the material-petrographic composition of coal beds. In Zone I they consist mostly of fusainized components, the products of a subaerobic alteration of plant material under slightly inundated conditions. Predominant in Zone II are preliminarily gelled, then slightly fusainized, plant remains whose accumulation took place in the unstable environment of abundant water, often flowing, with alternating aerobic and anaerobic conditions. Finally, the coals of Zone III are gelled to a considerable extent, having been formed under highly inundated conditions, mostly as a result of anaerobic processes.

Related to the marsh inundation factor is the type of coal-making vegetation: forest and forest-marsh in the southern part and quagmire in the north of the basin. It appears that the greater inundation and the generally lower relief in the north promoted the accumulation of allochthonous elements in coals (a higher content of ash, attritus, relatively stable plant remains, etc.). The appearance of such elements in coal beds, accompanied by a gradual reduction in the latters' number and thickness, is one of the indications that the basinward reduction in the coal content in the Middle Jurassic sediments is genetically related to their wedging out (facies change).

These changes in facies have been considered for the entire Middle Jurassic section of this basin as a whole. However, individual sections also exhibit facies zonations of their own, although not every section contains all three zones. For instance, while the middle interval does show all three zones, laterally, with Zone I the best developed, only the second and third zones occur in the upper section, with Zone III being the best developed.

We see than that the petrographic and facies composition of the coal beds in the Orsk basin are closely related to their thickness, their vertical and lateral position in the section, and the general formation pattern of the coal measures. The main reasons for the observed changes are the paleogeography and tectonics of the region. The paleogeographic environment has determined the position of the deposition area, the deposition medium, relief, and the distribution of facies. The lateral zonation of facies is a function of the distance from shore, on the one hand, and the result of a general lowering of relief, on the other.

Inasmuch as, in this instance, the distance increase from the source of sediments toward the north coincided with the general lowering of relief in the same direction, the conditions were favorable for a distinct areal zonation of sediments. The area appears to have subsided differentially. Judging from a number of features (the abundance of coal beds, some of them fairly thick; the predominance of fusains; the presence of alluvial deposits), a slow subsidence can be inferred for Zone I, with the top of the deposits often above water level. Conversely, for the second and particularly the third zone, the subsidence was comparatively rapid. Such a situation, combined with the greater inundation, resulted in a smaller number of coal seams, their slight thickness, and the predominance of gels. This differentiation of the tectonic regimen promoted an even more distinct zonation.

It should be noted, however, that deposition below the water level in Zones II and III, can be explained also by a reduced amount of clastic material brought in, because of the distance from the source; this situation, with the same subsidence rate as for Zone I, could have resulted in under-compensation. It appears that both these factors were active.

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Received, 27 January 1960

ALGAE AND THE DEPOSITION OF CARBONATES[1]

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Deposition of carbonates in an aqueous medium is a well-known process and is well covered in the literature. At the same time, a closer look reveals that this process is complex and varies considerably under different conditions. Carbonates can be deposited in the following six ways.

1. Mechanical settling of carbonate suspensions — in no way different from that of clastic sediments.

2. Chemical precipitation, well covered in the literature (N. M. Strakhov and others).

We are not concerned here with these two.

3. Organic precipitation of lime [19, 22, 24], associated with the life processes of animal and plants which fix the dissolved lime and secrete it as part of their bodies.

This method of lime precipitation is best known from chara algae which carbonatize only their oosporangia shell. We find a brief mention of this in H. Horn af Rantzien's work [19]. A mature oosporangium of chara algae begins to carbonatize only after fertilization of the oogonium which then is modified to an oospore. At the same time, the spiral cells of the outer oosporangium sheath secrete lime out of their cell fluid and on their inner walls facing the oospore and adjacent spiral cells. The outer cell wall of extant characeae is free of lime. C. Davis [16] found calcium in the cell fluid of modern chara algae, where it is present in considerable amounts as an organic compound. K. Collander [14] has shown that the calcium content in the cell fluid of these algae is higher than in the water of their habitat. Because of this "organic" precipitation process, the calcareous sheath of the chara algae oosporangium acquires a peculiar stratified structure in both the modern [4, 19, 23] and fossil [4-7, 15, 18] representatives of these plants. The X-ray crystallographic analysis has shown [19] that the mature calcareous sheaths of living Chara Globularis consist of calcite. According to the petrographic data of several authors, the extremely fine (2-3μ) pure calcite layers alternate with those enriched in organic matter; as a result, the inner layer of the calcareous sheath exhibits a fine striation.

It appears that a similar but less known carbonatization process of cell sheaths takes place with certain red algae and continues during their lifetime. As a result, the thallus of the red algae of the family Corallinaceae hardens; the lime preserves the living structure which then can be studied in fossil state. These are the so-called "lithothamnia" or "nullipora".

Calcareous cell sheaths of the lime secreting red algae (Corallinaceae) studied by G. Bass-Becking and E. Galliher [11], consist of a pectin substance, isotropic in a transverse section and birefringent in the longitudinal. They seem to consist of fibers arranged tangentially to the walls. Elongated tablets of the fibers alternate with concentrically arranged slits. Individual calcite crystals attain tenths of a micron in length. The two authors believe that the magnesium deposition in these algae is a secondary phenomenon.

According to Berthold's observations, the calcareous cell sheath is thicker in the better lighted specimens. Some red algae fix up to 36% magnesium carbonate; however, magnesium is quite scarce in fossils. This is because $MgCO_3$ is present in the cell sheaths as an isomorphic addition to $CaCO_3$ and is readily leached out in the fossilization process. Cell walls in some fossil genera of red algae show a wavy extinction under crossed Nicols which suggests a regular arrangement of fine carbonate crystals. Whether this feature is original with living algae or is a result of recrystallization — has not been solved.

Assigned to the "organic process of lime deposition" may be the limy secretions of

[1] Vodorosli i karbonatoosazhdeniye, (pp. 81-86).

microscopic algae — coccophorida. These are minute planktonic organisms which form a calcareous "armor" around their cell, consisting of assorted indivisible particles (coccoliths). The process of their formation is not sufficiently clear as yet, but their study with the electron microscope has shown that they consist of extremely small (microns and fractions of a micron) flat calcite crystals regularly arranged, usually radially or tile-like. Under crossed Nicols, well preserved coccoliths look like spherulites.

Thus, individual large groups of algae (as well as of animals) form microstructures of their own as they secrete "organic" lime. These structures afford means for a more detailed study of the fossil remains of these organisms, determining some forms down to species. The precipitation of calcium and magnesium, as a result of this process, does not depend on their amount dissolved in the water.

4. Another carbonate forming process is called the "physiological deposition of lime" [8, 22, 24], wherein irregularly arranged carbonate crystals are precipitated on the cell surface. This process is brought about by photosynthesis by the plant in a water medium. This process requires that the plant fix calcium bicarbonate of H_2CO_3, HCO_3^-, and CO_3^{2-} in the photosynthesis [9]. All of these components are absorbed by the lower surface of the leaf. The lime is secreted in its upper part as $Ca(OH)_2$; reacting with $Ca(HCO_3)_2$, the latter forms solid $CaCO_3$. Such a "physiological polarity" takes place in those plants which have a bilateral structure.

The chara algae have no such structure, their entire thallus being the assimilating organ consisting of a system of assorted rods without any morphological and physiological differentiation into leaves or the upper and lower surfaces of assimilating elements. However, the vegetative parts of characea secrete lime in just that way. K. Arens [10] has identified a "physiological multipolarity" with living Nitella flexilis, Chara baueri, Ch. braunii, and Tolypella nitricata, which was expressed in annular lime incrustations about the rods, secreted by the annular zones of cell sheaths. A portion of these zones fixed calcium hydroxide; the other, alternating with the first, secreted it. As a result, there is the continuous calcareous sheath covering all vegetative parts of the chara algae.

The structure of these sheaths is quite different from that of the oosporangia in the same plant, described above as the result of an "organic deposition of lime". The sheaths of the vegetative parts in chara algae are finely- to coarsely crystalline. It is of interest to note that certain fossil charophytae which form additional calcareous sheaths around their oosporangia by means of vegetative growths, also have a granular-crystalline structure.

A similar process takes place with green (siphonales) algae, where bare assimilating parts of the thallus fix calcium from the solution and deposit it, also as a sheath, in certain areas of the outer cell surface. These features are quite typical of siphonales; they make it possible to reconstruct nearly the entire plant from its fossil remains.

Lime secretion by certain green (Chaetophora) and blue-green algae, forming local sheaths, apparently belongs to the same group of processes. A. A. Yelenkin [2] describes Scytonema drilosiphon, a blue-green alga which he studied in cooperation with V. V. Polyanskiy. It grows largely in greenhouses (aerophytae); in a fern greenhouse, it covers sand shelves, tuffs, brickworks, etc. He writes, "In most instances, all filaments were covered with more or less well-formed calcareous sheaths, with only a few filaments showing very fine sheaths or none at all. These sheaths are often arranged along a filament in individual cylindrical sections, or as rings locally baring a mucous sheath... Of particular interest is the fact that these sheaths are formed on material free of calcareous salts such as sand and wood. This constitutes incontrovertible proof that Sc. drilosiphon, indeed, is a peculiar independent species having an inherited capacity of extracting calcium from a soil where it is barely present, and of using that calcium for dressing up its body." He adds [2, p. 383] that Sc. drilosiphon may take its calcium from irrigation water, but does not know the details of the process.

Judging from the structure of these annular sheaths, it may be supposed that the process of their secretion is similar to that for the calcareous sheath of the vegetative parts of chara algae. In blue-green and green algae, lime is secreted in the mucus which covers the thallus.

Thus, the difference between the "organic" and "physiological" deposition of lime is that in the first, lime is secreted from the cell fluid, within the cell or in its walls; in the second, the lime is precipitated on the outside of the cell walls. In the "organic process, the mineral precipitate consists of fine calcite crystals, regularly arranged to give definite optical effects (wavy extinction, etc.) when they are well preserved, and show an alternation of fine organic layers. The microscopic structure of calcite thus precipitated is different for different algal groups. In the "physiological" process, the calcite is usually free of organic matter, coarse-grained, the crystals not showing any regular extinction pattern. While the anatomy of the algal cells is

preserved in the "organic" process, the "physiological" process results in a calcareous sheath about the thallus; the latter's external form is fossilized, rather than that of the cell. It should be added that some fossils exhibit granulation, or a transition from the original calcite to a fine-grained pelitomorphous carbonate, as the result of the recrystallization of limy algal secretions.

5. The next process of carbonate deposition may be called "biochemical". Here aqueous plants play an indirect part by assimilating carbon and thus modifying the pH of water. Even here the carbonate deposition is accomplished in two ways: 1) the plants modify the water pH throughout the area of their distribution; as a result, the carbonate is precipitated in crystals evenly distributed over the bottom; and 2) lower algae (apparently along with bacteria) modify the pH only in the restricted habitat of a given colony. In this case carbonate crystals are precipitated locally only.

The first method was studied by L. Blanc and K. Molinier [13]. As anticipated, this carbonate is in no way different from the chemically precipitated, and usually presents a pelitomorphous carbonate sediment.

The second instance has not been adequately studied; however, some botanists [1] describe carbonate crystals formed among algal filaments as a result of the life activity of the algal colony. These crystals are quite similar to those chemically precipitated; however, being formed among the algal filaments, they may fill the entire space occupied by the algae and thus produce a structure different from that of the usual sediments.

V. O. Kalinenko's experiments [3] have shown that with a localized modification of pH caused by intensive bacterial activity, carbonate is precipitated as round granules in the bacteria colony zone. These data are corroborated by the work of C. Lalou [20] who experimented in a marine aquarium. The sediment he obtained as a result of thiorhodobacterial life activity was spherically granular, similar to that obtained by V. O. Kalinenko.

6. In a "mixed" or "stromatolithic" process, carbonates are deposited by the chemical, biochemical, and occasionally the physiological processes, with some mechanical deposition of terrigenous material, in various proportions. This instance is the least familiar, under the present conditions, with some light shed on it by the work of M. Black [12]. In past geologic periods, this "mixed" process was responsible for the peculiar and striking carbonate bodies — the so-called stromatoliths (often called "algae"). Seasonal changes in the living conditions of lower algae (changes in temperature and salinity) modified the number of individuals and species of algae, thus producing changes in the amount of calcium deposited and in the manner of its deposition. It is possible that when an alga dies, bacteria take up the work of carbonate precipitation on the stromatolith surfaces. This assumption is corroborated by the presence of rounded carbonate granules in carbonate sediments and in stromatoliths; these granules are often taken for algae and are similar to those obtained by V. O. Kalinenko [3] and C. Lalou [20] in experiments with bacterial colonies.

Stromatoliths are being formed at the present time (they are included in "tufas", in the broad sense), but they were particularly abundant in the past.

In addition to these, there are other, less common and virtually unstudied processes of carbonate deposition, such as the limy crusts in caves produced as a result of life activity of blue-green algae and bacteria in the absence of light [17, 21].

In evaluating the part of these processes in the past, it can be stated that late Precambrian (Proterozoic, Riphean, Sinian) witnessed the greatest intensity of the "mixed" and probably the "biochemical" processes of carbonate deposition. The "organic" and "physiological" processes emerged in the Cambrian, with the "mixed" process holding sway as late as the Silurian. Since then, it has been dying down gradually, persisting now to a very limited extent, while the "organic" process has been gaining in importance. That can be explained by 1) a change in the general living conditions of bottom organisms (not as many shallow epicontinental seas); 2) the greater numbers of organisms which have taken over the oceans as well as the shallow seas; and 3) the appearance of more complex organisms adapted to the variety of conditions and the carbonate contents in water.

We have seen that the "organic" and "physiological" processes are responsible also for structures visible only under the microscope. The "organic" process produces assorted calcareous organic remains identifying animal and plant groups. Obtained in the "physiological" process are carbonate crusts which allow a reconstruction of the external morphology of a plant — often in considerable detail. The "biochemical" and "mixed" processes leave behind bizarre and curious carbonate structures (either dolomitic or calcareous). The study of these fossil structures, initiated a century ago and still continuing, is fraught with obscurities and speculation. As yet, there has been no study made of these growths on living objects.

These processes of carbonate deposition are

interesting also in connection with their environment. While the "biochemical" and "mixed" processes are closely related to an adequate supply of dissolved carbonates, thus reflecting to a certain extent the carbonate salt content of a solution, the "organic" process is independent of its composition and concentration. The mere presence of calcium salts is enough. The "physiological" process seems to occupy an intermediate position, insofar as the intensity of carbonatization grows with the concentration of calcium salts; still, this is a process peculiar to certain algae and not to the others.

We see, then that the "organic" and "physiological" processes differ substantially from the "biochemical" and "mixed", the last two being more like the purely chemical precipitation of magnesium and calcium carbonates. It follows that Precambrian carbonate deposits containing organic structures of the stromatolith type are quite different from purely organic Cambrian and younger sediments and are closer to chemical sediments despite the presence of organisms. From the Cambrian on, the biochemical processes have been replaced, to an ever greater extent, by the purely organic and with a corresponding difference in the carbonate deposits.

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Received, 29 October 1960

AN EARLY PHASE OF THE DEVELOPMENT OF QUATERNARY MAMMALIAN FAUNA IN SOUTH EUROPEAN U.S.S.R.[1]

by

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A mammalian fauna designated by V. I. Gromov [8] as the Khaprovsk faunal complex was widely distributed throughout the U. S. S. R. in the early Eopleistocene, with horses, elephants, and assorted deer species as its principal elements. Their preponderance is obvious in any fossil site of that time.

The Khaprovsk faunal complex represents the fully developed fauna of a new type which gave birth to the principal groups of Anthropogene mammals. The assertion that elephants and true horses appeared here is not quite correct: these genera were already in ascendancy by that time. Their origin and the distribution of their early representatives undoubtedly dates back to an earlier time.

The Khaprovsk fauna in South European U. S. S. R. is more or less correlative with the Villafranchian of Western Europe. An ancient form of the southern elephant from those sites and from Africa has similar teeth structure to Archiskodon planifrons Falconer et Cautley, known from India. C. Arambourg [20] has designated the African form of this elephant as Archidiskodon africanavus. In Villafranchian fauna elephants similar to A. planifrons in teeth structure occur almost always together with Equus stenonis or E. Robustus.[2]

Both the Villafranchian fauna in Western Europe and the Khaprovsk fauna of European Russia are preceded by the Rousillion type of fauna of France. The latter (Tables 1 and 2), in turn, is still fairly closely related to the preceding Hipparion fauna, although containing the principal representatives of Anthropogene fauns (horses, deer of the genus Cervus, etc.). To be sure, the distribution of these new forms is far from complete; still, their presence suggests that the principal elements of this fauna originated in Roussillion time.

Mammalian bone remains of the Roussillion type of France have been identified from several localities in south European U. S. S. R. Only two sites are known from the Northern Caucasus — Kosyakino quarry (in the vicinity of Stavropol'-Caucasus) and the Dorurs quarry (near Armavir). In the Ukraine, a Russillion fauna has been identified in gravel deposits well exposed in the Kuchurgan River valley, near the villages of Novopetrovka and Trudomirovka.

Another Ukrainian site of a fauna similar to the Roussillion is located in the karst deposits of the "odessa Catacombs". Its composition (Anancus arvernensis, Paracamelus alexejevi, Ursus arvernensis, etc.)[3] suggests its similarity to both the Khaprovsk fauna and the Roussilion fauna of Moldavia. Its association with either one is a moot question for the time being.

In Moldavia, sites of the Roussillion type fauna [18, 19] are considerably more numerous although they are concentrated only in the southwestern part of the Republic: Gavanosy, Peleney-Moldavan, Fezeste-Moladavan, Novaya Karbolia, Budey, Musand, and Kislitsa. While working in southwestern Moldavia in 1959, together with K. V. Nikiforova, N. V. Rengarten, and N. A. Konstantinova, on exposures of a sand-gravel section along the high sides of the Bol'shoy Sal'ch and Kagul valleys, we collected the bone remains of turtles and mammals: Dolichopithecus cf. ruscinensis, Lepus sp., Ochotona sp., Spalax sp., Ochotona cf. antiqua, Ochotona ex gr. eximia-gigas, Sciuridae gen. indet., Rhinicerothidae gen. indet., Testudo sp., Clemmys sp, as well as fishes and birds. The presence in Moldavian Roussillion fauna of a large ape of the Dolichopithecus ruscinensis Depéret type, whose bones are fairly common in the Roussillion fauna of France, Romania, and Austria, was first established here.

[1] O ranney faze razvitiya chetvertichnoy fauny mlekopitayushchikh na territorii yuga yevropeyskoy chasti S.S.S.R., (pp. 87-96).

[2] The specific name, robustus, apparently is a synonym for Equus stenonis major.

[3] A full roster of mammalian fauna from the "Odessa Catacombs" is to be found in A. D. Roshchin's work [16, pp. 33-34).

Table 1

Russillion type fauna from the fossil sites south European U. S. S. R.

Fauna	Kosyakino quarry [1, 5, 6]	Rossillion fauna of Moldavia: Gavanosy, Musaid (19 & fr. author's material)	Kuchurgan gravel of Novopetrovka etc. [11]	Armavir (Dorurs quarry, fr. author's material)
Macaca sp.			*	
Dolichopithecus cf. ruscinensis		*		
Ursus cf. arvenensis	*			
Amphicyon (?) sp.	*			
Dinocyon cf. thenardi	*			
Vulpes vulpes fossilis		*		
V. sp.	*	*		
Felis cf. issiodorensis	*			
F. sp.			*	
Canis sp.	*		*	
Machairodus cultridens (?)		*		
Lynx brevirostris		*		
Hyaena sp.		*		
Hyaenarctos sp.	*			
Mastodon borsoni			*	*
Anancus arvernensis	*	*		*
Tapirus cf. arvernensis	*			
Dicerorhinus megarhinus		*	*	
D. orientalis	*			
Rhinoceros longirostris (?)			*	
Hipparion crassum		*		
H. sp.	*		*	*
Equus sp.		*		
Hippopotamus sp.		*		
Propotamochoerus provincialis	*	*		
Gazella sp.	*	*	*	
Procapreolus sp.	*			
Capreolus australis		*		
C. cusanus			*	
Procervus variabilis			*	
Pliocervus sp.	*			
Cervus (Rusa) moldavicus		*		
C. ramosus		*		
C. pyrenaicus		*		
C. cf. perrieri			*	
C. aff. pardinensis			*	
Pseudalces sp.	*			
Eustyloceros blanvilli			*	
Muntiacus flerovi			*	
Paracamelus cf. alexejevi		*		
Parabos boodon		*		
Sivatherium (?) sp.	*			*
Hystrix sp.		*		
Lepus lascarevi			*	
L. sp.	*	*		
Amblicastor cauvasicum	*			
Castor praefiber		*		
C. fiber			*	
Steneofiber sp.	*			
Cricetus sp.	*			
Mus sp.	*	*		
Ochotona ex gr. eximia		*	*	
O. cf. antiqua	*	*	*	
O. sp.		*		
Spalax sp.		*		
Prolagus sp.		*		
Sciurus (?) sp.		*		
Desmana sp.	*			
Sorex sp.	*			
Talpa sp.	*			
Crocidura sp.	*			
Erinaceus (?) sp.		*		
Clemmys pidoplickai			*	
Testudo sp.	*	*	*	
Clemmys sp.		*		

Table 2

Stratigraphic distribution of mammalian genera in the Anthropogene of south European U. S. S. R. from data of V. I. Gromov [8], I. G. Pidoplichko [13], I. I. Sokolov [17], A. K. Vekua [4], and from the author's material

Genera	Pliocene (Pontian and Cimmerian)	Anthropogene						Holocene
		Eopleistocene				Pleistocene		
		lower		middle	upper	lower	upper	
		Kosyakino fauna and Roussillion fauna of Moldavia[1]	Khapr. complex	Taman complex	Tiraspol' complex	Khozarian complex	Upper Paleolith complex	
Mastodon	· · · · ·	· · · · ·	· · · ·					
Anancus		———	———					
Archidiskodon			———	———	———			
Hesperoloxodon				?———	———?			
Mammuthus						———	———	———
Hipparion	· · · · ·	· ·	· · · ·					
Equus		———	———	———	———	———	———	———
Dicerorhinus	· · · · ·	· · · · ·	· · · ·	· · · ·	· · · ·			
Coelodonta						———	———	
Hippopotamus		———	———	———	———			
Elasmotherium			———	———	———	———		
Cervus		———	———	———	———	———	———	———
Alces			———	———	———	———	———	———
Pliocervus	· · · · ·	· · · · ·						
Megaloceros					———	———	———	———
Capreolus		———	———	———	———	———	———	———
Rangifer						———	———	———
Gazella	· · · · ·	· · · · ·	· · · ·	· · · ·	· · · ·	· · · · ·	· · · · ·	· · ·
Saiga						———	———	———
Capra					———	———	———	———
Ovis				– – –	———	———	———	———
Paracamelus	· · · · ·	· · · · ·	· · · ·	· ·				
Camelus						———	———	———
Propotamochoerus	· · · · ·	· · · · ·						
Sus	· · · · ·	· · · · ·	· · · ·	· · · ·	· · · ·	· · · · ·	· · · · ·	· · ·
Parabos	?	———						
Bison			———	———	———	———	———	———
Bos					?	———	———	———
Tapirus	· · · · ·	· · · · ·						
Felis	?	———	———	———	———	———	———	———
Ursus	· · · · ·	· · · ·	· · · ·	· · · ·	· · · ·	· · · · ·	· · · · ·	· · ·
Crocuta	· · · · ·	· · · · ·	· · · ·	· · · ·	· · · ·	· · · · ·	· · · · ·	· · ·
Hyaena			?———	———	———	———	———	———?
Gulo					———	———	———	———
Canis		– – – – –	———	———	———	———	———	———
Vulpes	· · · · ·	· · · · ·	· · · ·	· ·	· · · ·	· · · · ·	· · · · ·	· · ·
Agriotherium	· · · · ·	· · · · ·						
Amphicyon	· · · · ·	· · · · ·						
Meles	· · · · ·	· · · · ·	· · · ·	· · · ·	· · · ·	· · · · ·	· · · · ·	· · ·
Machairodus	· · · · ·	· · · · ·						
Macaca		– – – – –	———	———	———?			
Dolichopithecus		———						

. . . . Genera originated in the Miocene and earlier.

——————— Genera originated in the Anthropogene

[1] Moldavian complex (see "Voprosy geologii antropogene" ["Problems of Anthropogene Geology"]. Izd. Akad. Nauk S.S.S.R., 1961, p. 32).

The Roussillion fauna was widely distributed throughout Europe, in all probability at the very beginning of the Eopleistocene [12] immediately before the advent of the Villafranchian and Khaprovsk faunas. The European type of the Roussillion fauna of France is characterized by the following forms [23, 33]: Anancus arvernensis, Mastodon borsoni, Parabos boodon, Cervus ramosus, Propotamochoerus provincialis, Capreolus australis, Hipparion crassum, Tapirus arvernensis, Dicerohinus megarhinus, Felis arvernensis, Lynx brevirostris, and Vulpes donnezani. It has now been established that continental arenaceous-argillaceous deposits of Western Europe, containing a Roussillion fauna, are contemporaneous with the Mediterranean Astian stage [21, 22].

Roussillion fauna (Roussillion, Montpelier, Kosyakino, Gavanosy, Melusteni) is quite reminiscent of the Hipparion fauna, and it retains many typical forms of the latter (Hipparion, Agriotherium, etc.). However, unlike a true Hipparion fauna, the Roussillion complex displays a definite scarcity of genera. While each Hipparion fauna site yields two-three genera of Hipparion, as many of rhinoceros, several different species and genera of mastodons, assorted giraffe species (Paleotragus, Sivatherium, Eladotherium), etc., the Roussillion assemblage contains but a single Hipparion genus, two genera of mastodons, one of giraffe, and one (?) of rhinoceros. Occurring along with Hipparion fauna forms in the Roussillion assemblage, are such genera as Anancus, Cervus, Capreolus, Lepus, Trogontherium, etc., indicating this to be a new fauna. In analyzing the generic content of the Roussillion fauna of Moldavia (sites in the Kagul, Bol'shaya Sal'cha, and Prut River basins), Hungary (Barot), and Romania (Melusteni), I. G. Pidoplichko [14, p. 51] came to the conclusion that a large number of these genera could have been direct ancestors of some living species.

No bones of true horse, elephants, or oxen have been found as yet in the most typical Roussillion fauna sites of Europe (Montpelier, Roussillion, Perpignan). With respect to the points of origin of these groups, it can be said that there is evidence of such remains in deposits contemporaneous with the Astian, or nearly so.

Southeastern Asia is supposed to be the point of origin for the true elephants. The exact place and ancestry of the genus Archidiskodon, the oldest of the known elephants, has not been determined as yet. Only indirect data shed some light on the origin of this stratigraphically important group which spread over almost all of the world in a fairly short time.

It appears that elephants have their origin in some forms of the subfamily Stegodonthinae. Stegodons, as well as the associated Stegolophodons (apparently ancestors of the genus Stegodon), were distributed largely in the southeastern regions of the Asian sector of the Old World [31]. In Europe, their discoveries are quite rare [32]. There are no accurate data on their presence in the U. S. S. R., although there is a Stegodon cranium in the Zardabi Museum of Natural History (Azerbaidzhan), possibly from the Trans-Caucasus [2]. Most probably, true elephants originated somewhere in India or some other place in southeastern Asia, because the Pliocene stegodons of India and of Japan are closest to the ancient elephants in teeth structure. Specifically, Stegodon clifti and S. bombifrons, the most similar to the latter, are known from Pliocene deposits of India and Japan as correlative with the Upper Pontian, Plaisancian, and Astian [28, 36]. This does not necessarily mean, however, that elephants have descended from one of those species (possibly from a form similar to them).

In the Villafranchian, elephants of the Archidiskodon planifrons type are known to be distributed throughout the tropical latitudes of the Old World: Africa, South Europe, and southeastern Asia (India, Japan). It can be said that the Villafranchian was the heyday of elephants because they were already represented over immense regions of the Old World. It is, therefore, reasonable to assume that they had originated in an epoch immediately preceding the Lower Villafranchian, i. e., in the Plaisance-Astian.

The earliest discoveries of Archidiskon planifrons come from the Middle Pliocene of China [37]. The same deposits yielded the remains of Mastodonthidae and Stegodonthidae. That was the time of coexistence for the three groups of proboscidae (mastodons, stegodons, and elephants), lasting for quite a long period. Mastodons and stegodons persisted in the Villafranchian (although in much reduced numbers) when true elephants were holding sway.

Teeth of Archidiskon planifrons have also been found in Western Java, in the Kali-Glagan zone which is possibly correlative according to H. Movius [30] with the Tatrot zone of India.

H. Koenigswald [25], who studied these teeth, came to the conclusion that they belong to a form more primitive than the Indian Archidiskodon planifrons. Considering the rudimentary structure of the Javanese form of planifrons elephant, Koenigswald surmised that the Kali-Glagah deposits are older than the Pinjor of India which yields the bulk of Archidiskodon planifrons remains. With respect to the presence of A. planifrons teeth in the Tatrot zone of the Siwalik Range, D. Hooijer and E. Colbert [24] believe that Kali-Glagah deposits

respond to the Tatrot of India. It should be ed that M. Krishnan [10] thinks that it is ssible to correlate the Tatrot with the ian of Western Europe.

These discoveries of the genus Archidi-don elephants in the Kali-Glagah deposits of a and in the Tatrot of India indicate that the phants of Asia probably antedated those of rope. From their study of mammalian na from the middle and upper Siwalik zone, oijer and Colbert conclude that the Tatrot e is characterized by a fauna transitional m Pliocene to Pleistocene. Corresponding hat time in Europe (Tatrotian or Astian of European nomenclature) is a Roussillion e fauna, also transitional (although it does contain elephants).

A similar picture transpires from the study other typical Anthropogene forms (horses, r, oxen). In the Villafranchian, horses and e oxen lived everywhere in tropical latitudes he Old World.

Considering the fact that horses had entered ope from America [9], one might think ir first findings would be made in those as along the route of their migration. Up now, the oldest bone remains of horses have n found in India (Equus sivalensis [10],[4] m Upper Siwalik [Tatrot?] deposits of the th Himalaya). There is no unanimity of nion on the correlation of Siwalik deposits h its European equivalents (Pilgrim, Colbert, others). Recent works [10, 36] correlate Tatrot stage with the Astian or Plaisancian Western Europe (the 1949 and 1952 works of Denizot demonstrate that the Plaisancian is a facies of the Astian). This allows a gen-l correlation of the Tatrot mammalian na with that of the Roussillion complex of ance.

There are three references to discoveries Equus in the Roussillion fauna of Europe. first is by J. Simionescu [35] in his cription of the Bereşti site (Romania) fauna. be sure, his illustration of the masticatory face of a lower molar of horse raises bts as to that tooth coming with the Rous-ion fauna of that site. The second finding Equus sp. comes from the U. S. S. R., from Moldavian site of the Roussillion fauna. . Pidoplichenk's work [15] mentions the covery of Equus sp. and Cervus ramosus in Pliocene deposits near the village of ney-Moldavan (the Bol'shaya Sal'cha River in in southwestern Moldavia). Here, a d-gravel sequence containing a Roussillion type fauna rests on the Pontian and is overlain by loess-like loams. The presence of Equus sp. in the same deposits with Cervus ramosus — a typical representative of Roussillion fauna — suggests that representatives of genus Equus penetrated the present U. S. S. R. as early as the Astian. Considering that the Equus had originated in North America and then migrated to the Old World, it can be assumed that its migration route to Europe was by way of Asia (India) and then to some extent through the south of the U. S. S. R. It is therefore not surprising that the remains of its earliest representatives have been found in Asia (E. sivalensis of India), with rare findings in Eastern Europe: the south of the U. S. S. R. (Pileney-Moldavia), Romania (Bereşti), and Hungary (Barot). It is to be noted that the Equidae group (one-toe, in this instances), being one of the most fleet-footed, should have spread faster than elephants and true oxes.

Having entered the Old World with its favorable steppe conditions, the one-toe horse rapidly spread out. As early as the Villa-franchian, horses of the Equus stenonis type populated most of the Old World, with the possible exception of extreme northern Eurasia.

In many places, early Eopleistocene true horses still coexisted with hipparions, as witness the Khapry site in south European U. S. S. R. containing the fossils of both. The changing climatic conditions, accompanied by extensive development of a steppe landscape, were the main cause of changes in the pre-dominant equidae since the one-toe horses with their thick enamel hypsodont teeth were better adapted to dry steppes than were the three-toe hipparions whose teeth were strengthened as a rule, by the low crown and thin enamel which were more suited to succulant savannah grass.

Some of the later hipparions, persisting here and there, developed some "horse" features (a considerably lengthening of the protocone, a smoother enamel on the upper molars, etc.) associated with grazing on the rough steppe grass. Thus, L. K. Gabuniniya [7] describes a new species, Hipparion apscheronicum, from Upper Apsheronian deposits of the Caucasus (village of Shakhovo), which contains many such specialized features.

An ancient form of horse from south European U. S. S. R. is represented by two varieties of Equus stenonis (major and typicus), both present as early as the Khaprovsk fauna.

Another group, common in the Anthropogene, is the true oxen (Bison, Bos, Leptobos). Judging from recent data, India appears to have been the place of origin for the subfamily Bovinae [17].

The earliest Bos, Bison, and Leptobos are known from India and China. In China, Bison

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[4] D. Hooijer [24] believes that this discovery nes from the Pinjor.

sp. was found along with Archidiskodon cf. planifrons [27], in Pliocene deposits (tentatively, Middle Pliocene). In India, Bison and Bos [10] were identified in the Pinjor, while Leptobos [36][5] — as early as the Tatrot, correlative with the Mediterranean Astian stage, as noted before.

In the U. S. S. R., the oldest known bone remains of true oxen belong largely to the genus Bison. For instance, I. G. Pidoplichko [15] mentions the presence of Bison sp. in the Khaprovsk fauna of the Kaira village area. In addition, this author, in cooperation with I. M. Sukhov, collected a complete horn shank of a very short-horn Bison sp. (even as compared with Bison schoetensacki), from gravels near the village of Dolinskoye (Reninsk rayon, Ukrainian S. S. S. R.). In faunal content, these gravels, as well as those from the vicinity of Reni, may be regarded as correlative with the Khaprovsk sands of the Azov region.

Thus there are several Bison fossil sites in deposits correlative with the Villafranchian of Western Europe and North Africa.

V.I. Gromov's work [8] also notes the discovery of a Leptobos cranium in the Upper Pliocene gravels along the Psekups River (North Caucasus) which contains a Khaprovsk fauna. There is a communication by N.I. Burchak-Abramovich [3] on a discovery of Bubalus sp. in the Upper Pliocene of the North Caucasus. A study of the illustration presented by that author and of the material described (some of it exhibited in the Groznyy Petroleum Museum) suggests that these horn shanks belong to the genus Bison. Their slightly-developed keel-like ridges are common to the horn shanks of ancient bisons (such as the horn shank from the Dolianskoye area, and even some isolated specimens of the Bison schoetensacki horns from the Tiraspol' gravel; true keel-like ridges are typical of buffalos' horn shanks). Thus two genera of oxen — Leptobos (?) and Bison are present as early as the Khaprovsk complex of south European U. S. S. R.

As of now, there is not much known on the direct ancestors of the true oxen. It is believed [17] that Urmiobos, known from the Maraga site of Hipparion fauna, is the probable ancestor of Bos. Nor is it clear whether there is any genetic relationship between the genera Bison and Parabos, known from the Roussillion fauna of Europe. The horns of representatives of the genus Parabos, too, have three keel-like ridges on their shanks. It is quite possible that these are merely parallel rather than genetic features.

Our inadequate knowledge of ancient representatives of true oxen precludes the determination of the time and place of the origin of this group. The scattered data available suggest merely that these genera developed in the Astian, and that their migration route to Europe passed, in part, through south European U.S.S.R., as witness the discovery of Bison sp. in the Khaprovsk fauna.

Thus, it may be regarded as established that the early representatives of elephants, horses, and oxen (Leptobos) appeared in Roussillion time — along with the genera Cervus and Capreolus, common in the Anthropogene fauna.

The composition of Roussillion fauna (hippopothami, mastodons, apes, giant land turtles) undoubtedly suggests its thermophilic tendencies, as well as a fairly humid and hot climate. Incipient aridity, with the corresponding expansion of steppes, affected the composition of that fauna. Mastodons became scarcer, especially Mastodon borsoni. That species, common to the entire Roussillion faunal province, occurs only in Western Europe during the succeeding Khaprovsk time. In the U. S. S. R., isolated discoveries of M. borsoni within the Khaprovsk complex are known only from the extreme southwest. The stunting of mastodons was reflected also in their size. Thus Mastodon borsoni and Anancus arvernensis from the Roussillion complex always show larger teeth and bones than those of the Khaprovsk complex species.

In the transition from the Roussillion to the Khaprovsk fauna of the south Russian plain, apes, amphicyons, agriotheria, hippopothami, small hogs of the genus Propotamochoerus, and giraffes disappeared, and the tapirs became less common. Steppe- and steppe-forest forms were predominant over: horses, camels, elephants, oxen, and assorted rodents. Of course, that was not true steppe. The presence of deer (roe, elk, etc.), trogontherium (T. cuvieri), and other forms, suggests the presence of considerable forest area, most likely associated with river valleys.

Members of individual animal groups began to develop hypsodont molars, to accommodate themselves to the change from succulent plants of humid subtropical forests to the rough vegetation of the steppe. The animal kingdom changed with the plants. Some species died out, as mentioned before, or were confined to greatly reduced areas (the northern boundary of the distribution areas of Propotamochoerus, Hippopothamus, and Macaca receded far to the south). The animals left in those restricted

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[5] Hooijer [24] states that Leptobos is known in India only from the Pinjor.

as underwent changes to essentially new ns. For instance, the Anancus arvernensis n the Khaprovsk complex is different from giant Kosyakino A. arvernensis (smaller h and a somewhat different lower jaw — a rower symphysion, etc.).

At the same time, the generic composition he fauna was maintained. Many genera zella, Ursus, Meles, Crocuta), dating k to the Miocene, persisted throughout the hropogene (Table 2). The presence of these cene genera is in no way characteristic of Anthropogene fauna whose main feature is presence of new forms rather than the sistence of the old, often lingering through r last generations. Unlike the Khaprovsk a, the Roussillion complex contains the of the representatives of genera Agrio-ium, Amphicyon, Indarctos, Pliocervus; he same time, it contains, in the U. S. S. R., mber of new genera: Anancus, Equus, pes, Lynx, Trogontherium, Lepus, Cervus, Capreolus. These continue in the following provsk complex, with most of them further eloped throughout the Quaternary.

There are some references relative to gning the Roussillion fauna of Hungary to Quaternary in the M. Kretzoj monograph on the mammalian fauna and continental tigraphy of the Villanyi highlands in gary. The concluding chapter states that ot fauna should be regarded as the oldest al phase of the Quaternary. The Barot a composition is as follows [29]: Anancus ernensis, Mastodon borsoni, Tapirus garicus, Cervus pardinensis, Capreolus Ursus bockhi, Propotamochoerus pro-ialis, Prospalax sp.. Macaca sp., and us sp. In composition this fauna is ilar to the Roussillion as known from tpelier, Bereşti, Gődöllő, Tarbolia, etc. discovery of Equus sp., in the Barot fauna vell as in the Roussillion fauna of Romania south European U. S. S. R., confirms once e that true horses lived in Eastern Europe re they moved on to the west.

Thus the successive relationship between Roussillion (often called "Middle Pliocene") Khaprovsk faunas is unquestionable. As been shown, it was in the Roussillion fauna many leading Anthropogene genera made r appearance. Accordingly, it should be rded as the initial stage of this new period he Earth's development.

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Received, 29 June 1960

BRIEF COMMUNICATIONS

MASSIFS OF MINERALIZED SERPENTINE- AND PYROXENE ROCKS IN THE MANSK BELOGOR'YE SPURS (EAST SAYAN)[1]

by

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It is well known that the localization of basic and ultrabasic rocks is associated with sections of the earth's crust where folded structures are developed extensively, and are complicated by linear zones of deep-seated faults (Altay-Sayan, Salair, the Urals, etc.).

Such linear rifts are known from the junction area of the West and East Sayans (upper courses of the Kazyr, Kizir, Shinda, Sisim, and Balakhtison Rivers, on the one side; the Rivers Dzhebash, Kizhi-khem, Dotot, Kholug-Bash, Katun, Grishkina Rechka in the Kryzhin Range — on the other); associated with them are massifs of mineralized (titanium and iron), ultrabasic to basic rocks and syenites (Lysan group, upper courses of the Kazyra River and the Grishkina Rechka, and the Dorotsk, Gremyachinskoye ore deposits, etc.).

The work of many investigators of the West and East Sayans has resulted in the discovery of a sizable number of ultrabasic and basic intrusions having, however, widely diversified chemical features. The West Sayan ultrabasic and basic massifs contain numerous deposits of chrysotile-asbestos and chromite ores (Aktovraksk, the Izhim group, and other deposits). Many such massifs in the vicinity of Lysan Summit and in the junction area of the East and West Sayans contain titanium and iron mineralizations (the Lysan group; Upper Kazyr and Dotot mineralized massifs) without any chrysotile-asbestos and chromite deposits. Our studies covered the Lysan massif group consisting of the Lysan, Bezymyannyy, Piramida, Bol'shaya and Malaya Rossyp's, Skala, Kedranskiy, and Sisemskiy.

Involved in the geologic structure of Lysan Summit are metasedimentary Proterozoic rocks cut by ultrabasic to basic intrusions. The Lower Proterozoic sedimentary section is represented by the Derbinsk formation of marbles and marmorized limestones containing interbedded quartzites and shale beds. The thickness of the formation is about 2.5 km. This formation is overlain by amphibolitic and chloritic schists containing a variable amount of feldspars, sericite, biotite, epidote, and sphene as well as the amphibolites, effusives, and carbonate rocks of the Kuvay formation (Upper Proterozoic).

Exposed in the middle course of the Balakhtison River, on the east side of the ultrabasic intrusive massifs, are Lower Cambrian deposits which rest with a sharp angular unconformity directly on the Upper Proterozoic, and have a thick conglomerate at their base.

East Sayan is characterized by the presence of regional faults whose origin dates back to the Precambrian; new faults have been formed in later periods, particularly in the Ordovician.

Structurally, the Lysan area is the northwestern terminus of the southwestern limb of the East Sayan anticlinorium; the persistent zone of deep, regional, Shinndinsk-Derbinsk rift runs parallel to the latter.

Associated with this rift is the Lysan group of mineralized ultrabasic to basic massifs occurring in the Kuvay effusives and schists. The latter formation contacts the Derbinsk formations along the rift and forms steep northwest-trending faults, while the underlying Derbinsk metamorphics form gentle folds with a regional southwesterly dip of 60°.

The Lysan group of mineralized massifs forms a chain 25 km long, trending from northwest to southeast. The individual massifs are elliptic in plan, standing out sharply in an echelon arrangement typical of the entire area.

[1] Massivy orudenelykh serpentinitov i proksenitov v otrogakh Manskogo Belogor'ya Vostochnogo Sayana, (pp. 97-102).

The intrusive contacts dip 40-80° steeply the southwest, the average thickness of the ocontact zone is 2-2.5 m. The mineraliza- on is associated with basic and ultrabasic trusions (pyroxene- and serpentine rocks d gabbros) which occur together in each assif, conformably with their enclosing ıvay schists. These mineralized rocks have en traced by drilling to a considerable pth.

Serpentine rocks are not common among e mineralized rocks of the Lysan massifs. ıeir principal rock-forming mineral is rpentine (antigorite, chrysotile, less com-

Carbonatization of these serpentine rocks becomes more marked with depth. Present in carbonate veinlets are tabular crystals of ilmenite up to 10 mm long, also apatite, quartz, feldspar, chlorite, amphiboles, arsenopyrite, pyrite, and other sulfides. The serpentine rocks also contain isolated, fine grains of sericite, brucite, iddingsite, iron hydroxides, pyrite, chalcopyrite, pentlandite, and cobaltite.

Chemical analysis shows the following averaged composition of mineralized serpentine rocks, in % (I — fine-grained serpentine rock; II — same, coarse-grained, average of six analyses):

	I	II		I	II
SiO_2	24.33	19.69	K_2O	0.02	0.02
TiO_2	6.44	8.40	Na_2O	0.1	0.08
Al_2O_3	3.57	4.99	P	0.044	0.02
Fe_2O_3 (Total)	37.60	38.52	S	0.01	0.01
FeO	—	14.89	V_2O_5	0.04	0.14
MnO	0.24	0.31	Cr_2O_3	0.1	0.016
CaO	0.58	0.24	H_2O^-	0.02	0.05
MgO	23.39	21.90	Loss in heating	-up to 9.66	

only bastite). The serpentine content reaches %. There are serpophite veinlets up to 2 cm ick. The serpentine is developed on pyro- ne and olivine, whose fine relict grains have en preserved in the rock.

The second important serpentine rock- aking group is of ore minerals —titano- agnetite, ilmenite, and magnetite — whose ntent fluctuates in a broad range from 2 to % (18% on the average).

The usual proportion of ore minerals in the rpentinite, according to our data, is 80% for anomagnetite and 15% for free ilmenite.

The titanomagnetite is magnetite with nenite in tabular or punctate inclusions, or th. The punctate inclusions are concentrated ually in the periphery of titanomagnetite ains.

As a rule, the tabular ilmenite inclusions in anomagnetite constitute one tenth (up to 25%) the grain area, locally falling off to 3-5 tab- ts for a grain, one mm long. These tablets e usually 0.002-0.004 mm thick (up to 0.003 m), with the width-length ratio of 1/5 to 1/30. ıese ilmenite inclusions show extremely nuous outlines; they are often seen to corrode e titanomagnetite grains. Regular crystals of nenite, with rectangular outlines, are rare.

Magnetite is present in fine (0.2-0.5 mm) regular inclusions. The secondary minerals e chlorite, tremolite, talc, carbonates, and atite.

Mineralized pyroxenites are somewhat less common than the serpentinites.

All pyroxenites are altered to some extent: with their pyroxene uralitized, epidotized, and chloritized, and the ilmenite leucoxenized. These rocks consist mostly of pyroxenes (up to 80%) and ilmenite (up to 18%).

The pyroxenes are represented largely by monoclinic varieties (pigeonite and, to a lesser extent, by hedenbergite and diallage), less commonly by rhombic (hypersthene). Uralite is often developed on the pyroxenes, with chlorite and epidote developed on the periphery of their grains and in cleavage planes.

Apatite, sphene, biotite, and carbonates occur in isolated fine grains.

The ore grains are represented by ilmenite — 75-100%, and pyrrhotite — up to 20% of total ore minerals in the pyroxenite.

Titanomagnetite, magnetite, pyrite, chalcopyrite, and to a lesser degree, pentlandite, cobaltite, braviote, and violarite have been observed in isolated grains.

Ilmenite forms idiomorphic grains having simple, even, and rounded outlines (unlike the sinuous outlines of titanomagnetite grains), which indicates the almost simultaneous crystallization of ilmenite and pyroxene (with pyroxenite crystallized somewhat earlier).

The ilmenite often shows a polysynthetic

twinning. The size of its grains varies from 0.1 to 3.0 mm, being 0.8 mm on the average. The grains show a leucoxene fringe and occasional fine sphene grains. The ilmenite is locally replaced by leucoxene in typical lattice and skeletal structures.

Pyrrhotite occurs in inclusions with sinuous outlines and an average diameter of 0.2 mm (attaining 1.0 mm); these grains are often associated with veinlets (as are the other sulfides).

The chemical composition of the mineralized pyroxenites (average of six analyses) is as follows (in %):

SiO_2	34.2	K_2O	0.18
TiO_2	9.97	Na_2O	0.48
Al_2O_3	7.98	P	0.033
Fe_2O_3	4.70	S	0.27
FeO	14.76	V_2O_5	0.097
MnO	0.14	Cr_2O_3	0.002
CaO	14.82	H_2O^-	0.04
MgO	9.71	Loss in heating	-up to 2,65

The third variety of titaniferous rocks from the Lysan group of massifs is of gabbroids, locally changing to gabbro-amphibolites. The dispersion of minerals in the gabbroids is considerably lower than in the serpentine rocks and is of an accessory nature.

The mineral composition of some of the Lysan gabbroids is similar to that of uralitized pyroxenites, except that the gabbro shows a better development of completely sossuritized plagioclase.

The principal rock-forming mineral in the gabbros is hornblende of an uralite type. It is locally replaced by earthy masses of epidote and sphene, often by chlorite, less commonly by biotite. The interstices between large amphibole crystals are filled with a fine-grained aggregate of sossurite.

The amphibole content in gabbroids reaches 70-85% (transition to uralitized gabbro-pyroxenites); up to 25% albite, up to 40% epidote, 20% carbonates, and up to 10% sericite; in addition there are isolated grains of pyroxene (relicts), quartz, and accessory apatite and sphene.

The principal ore mineral in the gabbros is leucoxenized ilmenite followed by pyrrhotite and other sulfides. Titanomagnetite and magnetite occur, as a rule, in isolated grains.

The chemical composition of the mineralized gabbroids is as follows (in %, average of three analyses):

SiO_2	42.02	K_2O	0.42
TiO_2	4.28	Na_2O	1.53
Al_2O_3	14.58	P	0.02
Fe_2O_3	2.10	S	0.007
FeO	10.51	V_2O_5	0.04
MnO	0.12	Cr_2O_3	0.032
CaO	11.86	H_2O^-	0.16
MgO	8.15		

Titanium dioxide enters the composition of ilmenite, leucoxene, sphene, amphibole, and pyroxene, of gabbro rocks.

Thus, the mineralization of ultrabasic and basic Lysan massifs is represented by an uneven dispersion of titanomagnetite, ilmenite, magnetite, and sulfides — in serpentine-, pyroxene-, and gabbro rocks. The massifs differ in the content of valuable components. Thus, the poorest in serpentines — the Lysan massif — contains about 4.5% TiO_2 and about 20% Fe. The Skala and Malaya Rossyp' massifs are somewhat richer in titanium and iron. The richest — although uneven — mineralization has been observed in the Bol'shaya Rossyp' and Piramida massifs (average of 8.5% TiO_2, 28% Fe_{total}, and 0.1% V_2O_5).

It has been established as a general rule that the principal mineral in serpentinite is titanomagnetite, 60-95%; ilmenite, 10-15%; and magnetite, 5-10% — of the total ore minerals; sulfides occur sporadically and in isolated grains. In the pyroxenites, ilmenite accounts for 75-100% of the total ore minerals, the remainder being sulfides: pyrrhotite followed by pyrite, chalcopyrite, etc.; titanomagnetite and magnetite are present usually in isolated grains. In gabbro, the ilmenite content is, as a rule, 95-100% of the total ore minerals, the remainder being magnetite and titanomagnetite, with sulfides usually in isolated grains.

The ore mineral content fluctuates within a broad range: from 2 to 36% in serpentinites, with an average of 18%; and from 1 to 10%, an average of 4%, in gabbro; and 5-18%, an average of 9%, in pyroxenites.

We first noted the high TiO_2 content in the gabbros (7.34%) in 1956 in the Sisim massif (1.5 km southwest of the Rossyp' and Piramida massifs); it should be kept in mind, however, that part of the TiO_2 in the gabbro is associated with non-ore minerals (leucoxene, sphene, pyroxene, amphibole, etc.).

The recent work (1957-1958) of the East Sayan Expedition of the Krasnoyarsk Geological Administration has confirmed our conclusions on the possible existance of new areas of mineralized basic and ultrabasic rocks in the

isim massif. Two "ore bodies" have been dentified here (the "northwestern" and "southastern"), which consist of mineralized rocks imilar to those of the Lysan group. To be ure, the degree of their mineralization is inerior to that in the Piramida and Rossyp' rea, and is more like the Lysan massif minralization. The two holes drilled in the isim massif have traced the mineralized ocks down to a considerable depth.

The sequence of mineralization in the Lysan roup rocks is as follows: the first to be crysallized were olivine, pyroxene, and plagiolase — followed by magnetite and titanite, nd then by ilmenite; in pyroxene rocks, lmenite crystallized simultaneously with nagnetite, somewhat later than the pyroxene. he last crystallized were the amphiboles, erpentine, chlorite, epidote, carbonates, alc, and sulfides.

The average size of ore grains in dispersed res is 0.7 mm; the average thickness of the abular incrustations of ilmenite and titanonagnetite — 0.004 mm; and the diameter of he ilmenite particles in the titanomagnetite - up to 0.01 mm, seldom larger.

The ore fraction is dispersed, sideronitic, ith sinuous ore grains filling the interstices etween the silicates.

The average content of TiO_2 in the Lysan nineralized massifs is 7% (up to 10.5%) in erpentinites; 10% (up to 18.5%); in the pyroenites; about 5% (up to 7.4%) in gabbros; the orresponding figures for total iron are about 3% (up to 34%), in pyroxenites about 13% (up o 24%), and in gabbro about 10%. The ratio f TiO_2 to total iron is about 1/3 in serentinites, about 2/3 (1/1.5) in pyroxenites, nd 1/2 in gabbros.

The following classification of the minerlized rocks of this area may be made:

1. Fine-grained serpentinites, low in tanomagnetite; their composition (in %) is as bllows: TiO_2, 4.5-6.0; Fe_{total}, 18-24; V_2O_5, .03-0.1; SiO_2, 24.5; P, 0.4; and S, up to 0.05.

2. Fine- to coarse-grained, intermediate n their titanomagnetite content: TiO_2, 6-10.5; Fe_{total} 22-29 (seldom up to 34.23); V_2O_5, .03-0.15; SiO_2, 18.7-22.8; P, 0.01-0.07; nd S, up to 0.02.

3. Medium-grained mineralized pyroxenites, ow in ilmenite, restricted in their distribution: TiO_2, 5-8; Fe_{total}, 15-18; V_2O_5, 0.06-0.1; SiO_2, 35.3-38.9; P, 0.01-0.09; and S, 0.1.

4. Medium-grained mineralized pyroxenites, intermediate in their ilmenite content:[2] TiO_2, 8-18.5; Fe_{total}, 15-24; V_2O_5, 0.1; SiO_2, 33.9-34.5; P, 0.03; and S, 0.2-0.3.

5. Medium-grained gabbros, low in ilmenite: TiO_2, 4-6 (seldom up to 7.4); Fe_{total}, 8-15; V_2O_5, 0.04-0.06; SiO_2, 40.2-41.1; P, 0.02-0.4; and S, up to 0.01.

Differences between types 1 and 2 and between 3 and 4 are rather arbitrary, being largely those in their TiO_2 content.

Thus, these rock types are differentiated largely by their material composition.

The presence of nickel and cobalt has been established for the serpentinites; as determined by three analyses in the Chemical Laboratory of the All-Union Geological Institute (VSEGEI), it varies from 0.05 to 0.2% for nickel, and from 0.01 to 0.03% for cobalt.

Mineralization of basic and ultrabasic rocks in the Lysan massifs is of a magmatic origin with a subsequent autometamorphic phase (serpentinization, uralitization, sossuritization, the formation of gabbro-amphibolites and amphibolites, etc.).

The mineralization process falls into three stages: 1) the formation of deep-seated faults and the intrusion of ore-free gabbros; 2) the intrusion of mineralized gabbros and the splitting off of pyroxenites; and 3) the intrusion of olivine pyroxenites and, possibly, of peridotites.

In the subsequent process of autometamorphism all olivine pyroxenites (and possibly peridotites) were altered to serpentinites, with the gabbros locally altered to amphibolites.

The hydrothermal processes resulted in the formation of veins of carbonate, quartz, feldspathic, chloritic, and serpophitic content — sometimes with ilmenite, pyrrhotite, chalcopyrite, magnetite pyrite, arsenopyrite, actinolite, talc, biotite, epidote, garnet, and apatite.

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[2] We include as rich titanium ores those massive ores having a titanium content of up to 52% (ilmenite veins in the upper courses of the Tatarka River and in the Trans-Angara region); and up to 20-25% (titanomagnetites of South America and the U.S.).

Council for the Study of Productive Potential,
Academy of Sciences, of the U.S.S.R.,
Moscow

Received, 8 May 1960

LOSSES TO SCIENCE

by

V. V. Tikhomirov, and L. B. Bel'skaya

Ansel'm Frantsevich Mutul', Senior Scientist at the Institute of Geology and Industrial Minerals, Academy of Sciences of the Latvian Republic, passed away on August 23, 1959.

He was born on March 12, 1902, in Reseknensk Rayon of the Latvian S. S. R. and graduated from the Geological Institute at Strassburg (France) University in 1934. Upon his return to his country, he went to work for the Latvian Ministry of Finance. In 1938, he became a geologist at the Institute of Industrial Minerals. During the War, he joined the guerillas, was captured, and interned in a German concentration camp. He was liberated by Soviet armies in 1945 and returned to the Institute, which became a part of the Latvian Academy of Sciences in 1946.

A. F. Mutul' was an outstanding engineering geologist of Latvia; he worked out and introduced a method of hydrophobization of mineral components used in road construction; a number of his works deal with the hydrogeology and structure of various areas of the Latvian U. S. S. R. He is the author of over 100 publications, geological reports, papers, and recommendations. His engineering-geological map of Riga is still unfinished.

A. F. Mutul' taught soil science and conducted practical experiments in that field at the P. Stuchka State University of Latvia.

Nikolay Il'ich Sokolov, Senior Scientist at the Hydrogeological Laboratory, Academy of Sciences of the U. S. S. R., Doctor of Geologo-Mineralogical Sciences, passed away on October 15, 1960.

He was born in Gatchina, Leningradskaya Oblast' on July 16, 1907, and graduated from the Moscow Geological Exploration Institute in 1931. While a student, he began his research work under A. D. Arkhangel'skiy. He was Senior Scientist for the Institute of Geologic Sciences, Academy of Sciences of the U. S. S. R., and worked in the Caucasus, Crimea, Donbas, Samara Bend, Bol'shezemel'skaya Tundra, and the Angara-Yenisey and Noril'sk regions.

The field of his interest was wide. He is best known for his works in tectonics, geomorphology, Quaternary deposits, karst phenomena, paleontology, the geology of ore deposits, hydrogeology, and engineering geology. His study of permafrost and of sites of primitive man are of considerable interest. He proposed a statistical method of measuring the tectonic jointing of rocks; the formula came to be known as "Sokolov's formula".

Starting in 1930, N. I. Sokolov carried on extensive teaching work, first at the Soyuzstroy Hydrogeological Technicum, Moscow Geological Exploration Institute, and Moscow Mining Institute, then at the Noril'sk Mining and Metallurgical Institute; in 1954-1956 he was lecturer (docent) at the Irkutsk Mining and Metallurgical Institute.

He was also an active member of the U. S. S. R. Geographical Society and the Moscow Society of Nature Students.

Tat'yana Aleksandrovna Boldyreva, Senior Scientist at the Institute of Industrial Minerals, Academy of Sciences of the U. S. S. R., died on October 21, 1960. She was born on May 9, 1913 in Rostov-on-the-Don, and graduated in 1939 from the Geologic Exploration Department of the Far-Eastern Polytechnical Institute, Vladivostok.

For 20 years she worked in various regions of the Soviet Union, mostly on the study of coal deposits. Her most important achievement is in the petrography of coals from the L'vov-Volynsk basin, in which she was engaged since 1952. She also participated in compiling the Coal Atlas of that region. Of considerable practical importance is her discovery of the relationship between the technological properties of L'vov-Volynsk coals and their depositional conditions.

Fedor Ivanovich V'yunov, lecturer at Tomsk Polytechnical Institute, the Stalin Prize Laureate, Candidate of Geological and Mineral Sciences, Member of the CPSU, died on December 12, 1960.

F. I. V'yunov was born in Saratovskaya
blast' February 21, 1908; graduated from
he Geological Exploration Department of the
omsk Industrial Institute in 1938; and set out
o study the geology of the mineral deposits in
astern Kazakhstan. He participated in the
tudy of major deposits of polymetallic ores
n the Altay and directed the work of discover-
ng large mineral reserves for the mining
nterprises of that region. He was appointed
n 1957 as lecturer at the Tomsk Polytechnical
nstitute.

Oleg Dmitriyevich Levitskiy, Corresponding
lember of the Academy of Sciences of the
. S. S. R., Laureate of the Stalin Prize, and
ne of the leading Soviet geologists in the field
f endogenetic ores, passed away on January
4, 1961. His obituary is found in this journal,
lo. 6, 1961.

Vasiliy Maksimovich Ponomarev, Doctor of
eologic-Mineralogical Sciences, Member of
ie Communist Party, Senior Scientist at the
. A. Obruchev Permafrost Institute, Academy
f Sciences of the U. S. S. R., died February
, 1961.

Born on August 13, 1905 in Chernyy Yar,
talingradskaya Oblast', V. M. Ponomarev
raduated from the Geological Exploration De-
artment of the Moscow Mining Academy in
930 and set out to study the hydrogeological
onditions in the Kuznetsk coal basin.

In the 1934-1942 period he was in charge
f the Hydrogeological Service and the Division
f Geology of the Glavsevmorput' Mining
nd Geological Administration. He organized
network of special permafrost stations which
ave collected, with his active participation,
nany valuable data in that field.

In 1942-1946 he worked for the Noril'sk
lining and Metallurgical Combine studying
on-ore minerals; from 1946 on, he was with
ie Academy of Sciences U. S. S. R., first as
irector of the Anadyr and then of the Aldan
ermafrost stations. Of importance are his
orks on the formation of ground waters on
ie North Sea coasts in the permafrost zone.

V. M. Ponomarev was awarded a Certificate
f merit from the Glavsevmorput' in 1940; in
952 he was awarded the Presidium of the
cademy of Science Prize.

Stepan Dmitriyevich Batishchev-Tarasov,
aureate of the Lenin and Stalin Prizes, Corre-
ponding Member of the Academy of Sciences,
azakh S. S. R., Member of the Communist
arty, died on March 26, 1961. He was an
utstanding expert on the exploration of non-
errous metal ore deposits.

C. D. Batishchev-Tarasov was born on August 14, 1911 in Kuybyshevskaya Oblast'; graduated from the Geological Exploration Department of the Leningrad Institute of Mining in 1935. In 1937-1941 he studied the Orsk-Khalilov deposits of natural-alloy iron ores, and the deposits supplying reserves for the Karaganda Metallurgical Combine; later, he worked on the iron and nickel reserves in one of the Ural regions and became Chief of the Geology Division at the Uralchermetrazvedka Trust in 1944. He was awarded the Lenin Prize in 1957 for his discovery and exploration of the Sokolovsk-Sarbay group of deposits. He was awarded the Lenin Prize.

For the last 15 years he was busy on problems of the Trans-Ural region, persistently promoting that little known province as a potential source for a number of minerals. From 1953 on, he was also interested in the irrigation of virgin and fallow lands in Kazakhstan.

S. D. Batishchev-Tarasov was awarded several medals of the U. S. S. R.

Sergei Petrovich Rodionov, Corresponding Member of the Academy of Sciences of the Ukrainian S. S. R., Doctor of Geological and Mineral Sciences, Professor, Member of the Communist Party, died May 2, 1961. He was Chairman of the Division of Chemical and Geological Sciences at the Academy of Sciences of the Ukrainian Academy of Sciences.

He was born in Zagorsk, Moskovskaya Oblast', October 8, 1898. In 1929 he graduated from the Geological Exploration Division of the Dnepropetrovsk Mining Institute, and remained there to teach mineralogy and crystallography; at the same time he was a collaborator at the Ukrainian Section of the Geological Committee. He was Chief Engineer and Director of the Krivoy Rog Geological Exploration Base in 1931-1935, while holding the chair of geology and petrography at the Krivoy Rog Mining Institute.

Subsequently, S. P. Rodionov moved to Kiev and worked for the Ukrainian Geological Administration — from 1938 on, at the Institute of Geological Sciences, Academy of Sciences of the Ukrainian S. S. R., in charge of its Scientific Division, then as its Acting Director, and as Director of the Geological Museum. In 1945-1952, he held the chair of mineralogy and crystallography at Kiev State University.

The principal publications by S. P. Rodionov deal with the petrography of the Ukraine. He studied the weathering crust and Precambrian geology of the Ukrainian crystalline shield, also the geology and petrography of Krivoy Rog. Of great scientific interest are his studies of the Teterev-Bug series of the Upper Archaean metamorphics in the Krivoy Rog region, also his study of meteorites.

S. P. Rodionov actively participated in the work of the Commission on the Geological Investigation of the U. S. S. R. and was acting Chief Editor of the "Geologicheskiy Zhurnal" (Geological Journal) of the Academy of Sciences of the Ukranian S. S. R. He is the author of over 60 monographs and articles, including several on the history of geology and a number of popular science works.

S. P. Rodionov participated in the Civil and Great Patriotic Wars. He was awarded the orders of the Red Star and of the Toilers' Red Banner, and medals of the Soviet Union.

Igor' Petrovich Tikhonenkov, Candidate of Geological and Mineral Sciences, Senior Scientist of the Institute of Mineralogy, Geochemistry, and Crystallography of Rare Elements of the Academy of Sciences of the U. S. S. R., Member of the Communist Party, died on May 21, 1961.

He was born in Novocherkassk, January 25, 1927; graduated from the Moscow Geological Exploration Institute in 1950; and worked at the Institute of Geological Sciences of the Academy of Sciences of the U. S. S. R., on pegmatites and rare-earth minerals in the various alkali massifs of the Kola Peninsula, Yenisey Range, East Sayans, Tuva, and the Kuznetsk Alatau.

I. P. Tikhonenkov was particularly interested in the distribution pattern and migration of certain rare elements. As a result of his studies of the Khibino massif, he developed a new approach to the origin of its inner ring zone and demonstrated the metasomatic effect of post-magmatic solutions — a fact important theoretically and practically.

An outstanding expert on the geology of aluminum ores, Yuriy Konstantinovich Goretskiy, Doctor of Geological and Mineralogical Sciences, Member of the Communist Party, died on an expedition, June 27, 1961.

He was born in Leningrad, January 27, 1912; graduated from the Moscow Geological Exploration Institute in 1935, and worked ever since at the All-Union Institute of Mineral Raw Materials of the Ministry of Geology and Mineral Conservation of the U. S. S. R. Here he was in charge of the study of the material composition of various aluminum ores, of determining the principal prospecting criteria, and of working out a classification of bauxite deposits.

Yu. K. Goretskiy was well known for his studies of sedimentary lithology. His early works dealt with the mineralogy and petrography of siliceous rocks, their origin and classification; and with the study of deposits and formation conditions of refractory clays — kaolinolites. From 1941 on, he was engaged in the search for and study of bauxites. He was credited with discovering the Salair bauxite region. In 1954 he compiled a bauxite map of the eastern regions of the U. S. S. R.

He was a member of the Geological Experts' Council at the Ministry of Geology and Mineral Conservation and of a number of commissions under the Division of Geologic and Geographic Sciences, Academy of Sciences of the U. S. S. R.

Yu. K. Goretskiy was awarded the Order of Merit and several medals.

Grigoriy Ivanovich Grachev, Candidate of Geological Mineralogical Sciences, Senior Scientist at the All-Union Scientific-Research Petroleum Geological Exploration Institute (VNIGNI), Member of the Communist Party, died in the performance of duty, July 12, 1961.

He was born in Ryazanskaya Oblast', January 1, 1911. In 1935 he graduated from the Geological Exploration Department of the Moscow Petroleum Institute and then worked in the oil regions of Central Asia as a party chief, senior geologist for "Nefterazvedka", and acting Chief Geologist of that Trust. He participated in the discovery of several new oil fields in the Fergana Valley and South Uzbekistan.

He served in the Soviet Army, from 1941 to 1946; after demobilization, he occupied a number of responsible positions in the Moscow Affiliate of the VNIGRI, and in head offices of Glavburneft' and Glavnefterazvedka, of the former Ministry of the Petroleum Industry.

His scientific works deal mostly with the history of geological development and the oil and gas potential in the Tadzhik depression and in the Tertiary Fergana trough, also with the Jurassic, Cretaceous, and Quaternary of those regions.

G. I. Grachev was awarded the Order of the Red Star and several medals of the Soviet Union.

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REVIEWS AND DISCUSSIONS

ON THE SKARN DEPOSITS IN THE CENTRAL AND NORTHERN URALS[1,2]

by

V. I. Smirnov

Skarns or as they are sometimes called, contact-metasomatics, are very interesting endogenetic formations, generally and especially those from the Urals, studies of which have disclosed many aspects of deep-seated magmatic processes. However, the difficulty of studying them has been responsible for the scarcity of major monographs on this subject, both in this country and abroad. This is why it is a pleasure to note the publication of a number of books on the contact-metasomatic deposits of the Central and Northern Urals, by the Mining Geology Institute of the Uralian Affiliate of the Academy of Sciences of the U. S. S. R. in Sverdlovsk.

We refer here to the three interesting books: 1) Ya. P. Baklayev, "Geologic structure and potential of the Tur'insk contact-metasomatic copper deposits in the Northern Urals", 1959; 2) V. A. Dunayev, "A mineral-petrographic description of the Techen deposit", 1959; and 3) L. N. Ovchinnikov, "Contact-metasomatic deposits in the Central and Northern Urals", 1960. This is a review of the latter, the most fundamental of the three.

More than 200 skarn deposits — mostly iron and copper, and, to a lesser degree, of other ores — are known to exist in the Urals. The history of their study and industrial development dates back to 1696 when cast iron was produced from Mt. Vysokaya skarn ores. The 250th anniversary of the Uralian ferrous metallurgy was celebrated in 1951. During some of the pre-Revolutionary periods, Uralian copper skarn ores produced up to 30% of total Russian copper. They attracted the attention of such leading geologists of the past as K. I. Bogdanovich, A. K. Boldyrev, N. K. Vysotskiy, L. Dupark, P. Ye. Yeremev, A. N. Zavaritskiy, A. P. Karpinskiy, N. Koshkarov, F. Yu. Levinson-Lessing, P. Palass, F. Poshepnyy, A. Stikney, Ye. S. Fedorov, G. Ye. Shurovskiy, N. N. Yakovlev, and other authors of works on the geology and mineralogy of the Ural skarns.

The Ovchinnikov book being reviewed treats the most important aspects of the formation of contact-metasomatic deposits for the Central and Northern Urals only; however, the material presented, as well as the masterly analysis and the important theoretical conclusions, make it of more than local interest.

Discussed in this monograph are the following aspects of the geology, mineralogy, and geochemistry of the skarns in the northern half of the Urals: 1) the tectonic pattern of their distribution; 2) skarn-forming igneous rocks; 3) types of skarn deposits and their characteristic; 4) types of skarns and skarn zones; 5) types of skarn ores; 6) mineral composition of skarn deposits; 7) additive elements in skarns; 8) the formation conditions of deposits; and 9) geologic exploration criteria.

The skarn deposits described show a definite zonal distribution: they are associated with the series of basic rocks, and their acid to alkaline derivatives, which intruded the Siluro-Devonian along the margins of geosynclinal troughs subsequently metamorphosed to greenstone synclinoria. Accordingly, there are two principal belts of skarn deposits: the main western, displaced somewhat east of the eastern gabbro-peridotite zone and toward the interior of the East Ural slope greenstone synclinorium; and the main eastern, also somewhat shifted toward the interior of that synclinorium, toward the western belt. In addition, there are four auxiliary belts associated with chains of gabbro-periodtite intrusions west, and particularly

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[1] O skarnovykh mestorozhdeniyakh srednego i severnogo Urala, (pp. 106-109).

[2] Review of L. N. Ovchinnikov' "Contact-Metasomatic Deposits of the Central and Northern Urals". Trudy Gorno-geol. inst. Ural'sk. filiala Akad. Nauk S.S.S.R., vyp. 39, 1960, 495 pp.

east, of the principal greenstone synclinorium. This zonal arrangement of skarn deposits is determined by deep faults which cut off the limbs of these greenstone synclinal troughs and controlled the intrusion paths of the basic magma to whose plagio-granitic and syenite differentiates the skarns are related.

The author thoroughly studied the distribution pattern of the skarn deposit zones with reference to the gabbro-peridotite formations as a whole, and particularly to the individual massifs and their outlines. He also studied the association of vein facies in the intrusions and skarns and the evidence of any geochemical relationship. As a result, he came to the conclusion that not all granitoids rocks are productive of skarn deposits but only those granitoids which are differentiates of a gabbro magma.

The book presents a detailed petrological description of hypabyssal intrusive massifs of a granodiorite and syenite offshoot originating from the same gabbro magma. They are characterized by the stable coexistence of orthoclase with hornblende and plagioclase of a higher basicity indicative of the greater activity of potassium in the solution. According to the author, this peculiar regimen of alkalis in skarn-forming granitoids may reflect their gabbroid origin as well as an assimilation of volcanic greenstone rocks by the magmatic melt.

L. N. Ovchinnikov differentiates all skarn deposits of this subprovince into four types according to their geology and morphology.

1. Bedded deposits in stratified volcanic-sedimentary strata: a) massive ores formed in the replacement of limestones; b) dispersed ores formed in the replacement of volcanic rocks.

2. Deposits at direct contacts of the intrusives and limestones and, less commonly, with other rocks.

3. Deposits in xenoliths of roof rocks among the intrusives.

4. Deposits in tectonic zones: a) in sedimentary-volcanic rocks; b) among the intrusives.

This approach — formalistic at first glance — makes it possible to determine many important aspects of the origin of skarn ore, including the part played by the stratigraphic, structural, and lithologic factors involved in the localization of the skarns. It has been established that rich magnetite ores are formed during the direct replacement of limestones, while only lean ores are formed in a superimposed mineralization on the skarns. Many deposits of the bedded type exhibit a regular decrease in the intensity of skarning and mineralization away from the intrusions, with a well-expressed zonation in the transition from massive ores to barren skarns and to unreplaced limestone (Mt. Vysokaya, Pokrovskoye, the Second and Third North Mines, etc.).

L. N. Ovchinnikov regards the skarns as the result of a chemical action of three media: the two contacting solid rocks and the solution saturating them in the combined processes of infiltration and the diffusion affected by the differential mobility of chemical elements.

Mineral parageneses of skarns and skarn zones are described separately for contacts of granitoids, syenites, and basic and ultrabasic rocks. Three types of skarn zonation are designated, along with a discussion of mineral associations for the individual zones, with consideration given to diagrams of parageneses as affected by the ratios of their aluminum, silicon, and calcium content.

The result of a consecutive superposition of mineral associations reflecting the progressively lower temperatures of a retrograde skarn-forming process, and the permanent reworking of mineral complexes, is considered with respect to D. S. Korzhinskiy's temperature stages and degrees of equilibrium.

Skarn deposits in the Central and Northern Urals contain major deposits of iron ores and, to a lesser degree smaller deposits of copper ores, even smaller concentrations of manganese ores, and shows of cobalt and molybdenum mineralization. According to the composition of their principal ore minerals, the iron and copper ores are differentiated into oxides (hematite and magnetite) and sulfides (Chalcopyrite-pyrrhotite, pyrite-chalcopyrite, and sphalerite-galena). The accumulation of ore minerals in skarns is closely related to skarn formation. Magnetite mineralization, beginning with the first step of equilibrium follows immediately after the formation of the primary skarn zones and lasts for a long time until the formation of low-temperature aposkarn parageneses in some deposits. The sulfide mineralization is definitely younger, being superimposed on magnetite ores, skarns, and aposkarn mineral associations.

Much space is given to the composition, manner of differentiation, qualitative distribution, the relationship and transformation of minerals in the skarn deposits, and to ore accumulations. The most common minerals are grouped as follows.

I. Minerals of Skarns And of Skarn Contact Rocks

1. Those occurring in all types of skarns:

a) anhydrous — garnets, pyroxenes; b) hydrous — vesuvianite, epidote, hornblende, ilvaite.

2. In skarn zones of syenite contacts: orthoclase feldspar, scapolite.

3. In magnesian skarns: phlogopite, olivines, spinel, cordierite.

4. In manganese skarns: bustamite, rhodonite.

II. Minerals of Oxide Ores

Magnetite, hematite, ilmenite, apatite.

III. Minerals of Aposkarn Associations

1. Those occurring in all types of skarns: actinolite, epidote, carbonates, quartz, prehnite, ilvaite, pumpellinite, chlorites, talc, zeolites, biotite, daphnite, and sphene.

2. In magnesian skarns: greenalite, tyn'ite.

IV. Minerals of Sulfide Ores

V. Minerals of the Oxidation Zone

A total of 32 skarn and 56 aposkarn minerals are described, differentiated into principal, subordinate, and rare skarn-forming compounds.

Thermo-analytical studies of hydrous silicates from the Ural skarns show that the sequence of their formation is accompanied by a regular increase in their water content and in a decrease in the dehydration temperature; accordingly, high-, intermediate-, and low-temperature groups of hydrous minerals have been differentiated.

A study of the distribution of the additive elements in the Ural skarns has revealed a number of interesting relationships. It has been established that the earlier generations of ore-forming minerals contain a larger amount of these additives. Their content decreases away from the massive ores toward the dispersed and from the central to the peripheral parts of the ore bodies. Their distribution is also more even in high-temperature associations than in the low-temperature. This is because the solubility of these additive elements and their mobility in infiltration and diffusion increase with the temperature. At the same time, because of the low concentration, their mobility in metasomatic processes is always lower than that of the principal ore-making elements. The latter's content changes gradually in diffusion, and abruptly, in infiltration. This feature affords a means of telling the diffusion metasomatism products from those of the infiltration metasomatism — by the manner of their distribution.

Thermodynamic conditions of skarn formation in the Northern and Central Urals are determined by their hypabyssal depths of 1200-1500 m, but not over 2000 m, at temperatures of about 200-800° C, and pressures of 100-500 atm. The book presents the results of many original experiments which demonstrate the many possibilities of isolating iron from a magmatic melt as the latter assimilates carbonate-bearing rocks.

On the basis of this occurrence of Central- and North-Ural skarn submeridional belts, the author has formulated prospecting criteria for skarn deposits and has charted prospective trends for the future.

A discerning reader will note the definitely "material", i. e., petrographic-mineral-chemical slant of L. N. Ovchinnikov's descriptions, with the spatial aspect of skarn-forming, the geologic aspect of skarn structure, their classification and characteristics being left in the background. However, even the most demanding reader will not fail to see the great theoretical and practical value of this monograph, which undoubtedly puts it among the most informative recent publications.

The book is illustrated by numerous tables, diagrams, geological sketches, and photographs, which are valuable in themselves. In this connection, it is regrettable that the quality of their reproduction, especially of the photographs, is poor. The indistinct macro- and microphotographs exasperate the reader; they must be printer's rejects, and the publisher should be sued.

L. N. Ovchinnikov studied the Ural skarns from 1940 to 1960, with time out for the War. After these 20 years, he contributed to the science of geology a major monograph which will take its place among other substantial works in the field of ore deposits. This is evidence that the time is ripe for reducing the voluminous field material on ore deposits of this country to a series of fundamental works on the individual groups, deposits, and metallogenic provinces of the Soviet Union.

Received, 26 October 1960

REVIEW OF M. F. NEYBURG'S BOOK, "CORMOPHYTIC BRYOPHYTES FROM THE PERMIAN DEPOSITS OF THE ANGARIDES"[3,4]

by

P. A. Mchedlishvili

The works of M. F. Neyburg have been conspicuous among our paleobotanical studies. In recent years she has authored a number of major publications on Paleozoic and Lower Mesozoic funas, with the book being reveiwed being of particular importance. Judging from the numerous notices coming from Soviet and foreign scientists, this book is regarded as an outstanding contribution to paleobotany. In addition, this is the first practical application of these fossil plants to the stratigraphic study of the Permian deposits of the Angarides.

Discoveries of pre-Tertiary bryophytes are quite rare; their Paleozoic remains, of indifferent value, have been known only from the Euro-American Paleozoic flora, and none from our own flora.

In 1941-1942 M. F. Neyburg first noted these bryophytes in the Upper Permian deposits of the Kuznetsk basin. In the following years she had at her disposal ample material on them from the Tungussa and Pechora basins, including the Lower Permian species.

She studied this material not only by the comparative-morphological method but mainly anatomically, using the latest paleobotanical techniques. Through her detailed study of Upper Paleozoic bryophytes, which were definitely associated with geologic sections, she turned a new leaf in the history of the bryophytes and, consequently, in the solution of problems of phytostratigraphy, paleoflora, and paleogeography.

This book has eight chapters, a foreword, and bibilography.

Chapter I discusses the history of the study of Paleozoic cormophytic bryophytes on the basis of a critical analysis of material from individual floral provinces.

Especially thorough is the author's treatment of fossil mosses from the Angarides, i. e., the Tungussa floral province where they were first found. The flora of that province have been studied by many others, including M. D. Zalesskiy; however, no one else has established the presence of cormophytic bryophytes here. The author notes that their remains were indeed encountered earlier in the Tungussa fauna, but that they were either overlooked or erroneously assigned — after a superficial study — to other plant groups, such as lycopods, equisetum, and conifers.

Chapter II and III discuss the stratigraphic basis of the distribution of fossil mosses, their occurrence and the state of preservation.

Of importance is the reference to the fact that mosses, when associated with coal measures, occur mostly in the argillaceous intercalations between coal beds and at their top — in grey, dark-grey, and black shales. This association will assist in the discovery of new fossil moss sites — and will serve as a criterion in the search for coal under the Angarides conditions.

Chapter IV discusses the methods used to study the material. Because of the various states of preservations of the moss remains, different methods were used to study them microscopically, such as maceration in various reagents, thin sections, transfer of fossilized leaves from the rock to a slide by means of Darras' Solution, still little used in this country, etc. This extensive and painstaking work, coupled with a detailed description of methods used in the study of fossil mosses undoubtedly will facilitate and broaden their use by paleobotanists.

M. F. Neyburg's work has always been distinguished by the amount of detail. This is especially true of this work where the treatment of plant remains is brought to perfection by a parallel consideration of their anatomy and morphology. M. F. Neyburg succeeded in synthesizing the various study methods, to revive the field of fossil mosses — so to speak — and to bring out their biostratigraphic value. The results of this synthesis of mosses are well illustrated in the remaining chapters of her book.

In Chapter V the author presents the basis of her systematic grouping of the plant remains and their taxonomic nomenclature. Her analysis of their principal characteristics shows that there is no essential difference between these and modern mosses except for certain anatomical and morphological features. Two principal groups of mosses have been identified using the entire community of their characteristics: subclass Bryales or green mosses and subclass Sphagnales or peat mosses. According to a number of their characteristic features, the latter are assigned to a new order, Protosphagnales ordo nov.

Despite its extreme complexity, the problem

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[3] O knige M. F. Neyburg "Listostebel'nyye Mkhi iz Permskikh Otlozheniy Angaridy", (pp. 109-111).

[4] Trudy Geologicheskogo instituta Akad. Nauk S.S.S.R., vyp. 19, 1960. Izd-vo Akad. Nauk S.S.S.R., p. 1-104. 52 illus. in text, and atlas tables I - LXXVIII.

taxonomic nomenclature has been solved ccessfully and correctly. As is well known, mains of fossil mosses not correlative with e modern species are usually assigned to a tch-all order, Muscites Brng. The author andons this formalistic approach and underores the considerable diversity of Paleozoic osses and introduces new generic names for em.

The differentiation of mosses and their stribution in the stratigraphic section are scussed in Chapter VI.

The author identifies 14 species of Paleozoic rmophytic bryophytes belonging to nine nera. Of these, 11 belong to the subclass yales and three (of the order Protosphagnales) subclass Sphagnales. All species are assign to their proper stratigraphic intervals, as ustrated with tables for the Kuznetsk and chora basins. These data, combined with ose for the Tungussa basin clearly demonrate that fossil mosses may be regarded as plant group supplementary — perhaps even cisive — in problems of stratigraphy.

Paleozoic mosses are described in Chapter I.

Subclass Bryales comprises the genera Intia species), Salairia (one), Uskatis (one), lyssaievia (2 species), Bajdaievia (one), chtia (one), and Muscites (one). With the ception of the last named, all of these genera ve first been described by M. F. Neyburg. e assigns three monotype genera to the order otosphagnales: Junjagia, Vorcutannularia, d Protosphagnum; the first and the third, too, ve been described by her first. The second nus was described by V. V. Pogorevich, but regarded it as a species of Equisetum.

This chapter presents a detailed charactertic of genera and species. This however, is t the usual formal morphological description; spite the length of the chapter, it does not ntain anything superfluous. The author's task the best possible representation of fossil mosses tainable through morphological and botanical udy. As a result, all of these extinct genera d species become biological rather than tificial entities, subject to natural classifiation. On the whole, this chapter, supplemented by good illustrations of the morphology d anatomy of mosses, is an excellent example of how to study fossil plant remains, to aluate their characteristic features to the inutest detail, and to reconstruct them as ologic objects. This portion of M. F. Neyurg's works is an object lesson that the escription of fossil plants is by no means a rmal art, like iconography.

Chapter VIII presents general conclusions. e shall pause here only for the principal ones which are particularly interesting and valuable.

While on the subject of Permian mosses, M. F. Neyburg discusses certain data on their evolution and phylogeny. Contrary to the prevailing opinion, she believes that the origin of sphagnum mosses should not be associated with the liverworts Jungermaniales, because Permian mosses do not show a close genetic relationship with the Permo-Carboniferous liverworts.

The author outlines the evolution of carmophytic bryophytes as follows: Intia — Protosphagnum — Lower Jurassic Sphagnum — modern Sphagnum. She admits that the evolution of the other branch of mosses — Junjagia and Vorcutannularia — is not as clear.

Material from the principal Angarides basins indicates a difference in the moss assemblages for the Lower and Upper Permian; at the same time it shows a similarity not only in genera but in species and their assemblages, and in individual intervals of the section. Therein lies the stratigraphic value of the mosses.

In discussing the relation of Permian mosses to paleogeography, M. F. Neyburg considers their general similarity to modern forms. From the ecology and distribution of the latter she infers that Permian mosses, too, inhabited largely temperate to cool and fairly humid zones. She comes to the conclusion that the Tungussa flora existed in a temperate climate and speaks of the relationship between cordaites and mosses, and outlines the contemporaneous phyto-landscapes.

Of interest are the author's explanations of the almost complete absence of moss remains among the Carboniferous and Permian flora of the Euro-American province: here, in tropical to subtropical climates, the mosses probably inhabited mountain regions and their remains never reached the lowland burial places. In the absence of mosses, peat must have accumulated here in marshes and peatbogs different from those of the Tungussa province.

M. F. Neyburg's conclusions are of great interest. Some of them are quite new and original with her; the others are at variance with the prevailing opinion but are always well substantiated by her careful analysis of most interesting new data.

The value of this book goes beyond the purely regional. It solves some problems and poses new ones. It will be of great theoretical and practical help to paleobotanists as well as to paleozoologist, geologists, botanists, and all natural scientists. Like the other works of M. F. Neyburg, this one also is of great

methodological value since it charts the course of work for young paleobotanists.

In the field of fossil mosses, this work has no peer in the world literature. It marks a new stage in the study of fossil floras; its scope and depth make it an outstanding contribution to domestic and world paleobotany.

Received, 22 March 1961

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M. F. Neyburg's "Cormophytic Bryophytes From the Permian Deposits of the Angarides" has brought forth much comment from home and abroad. For instance, Academician V. N. Sukhachev writes, "This work is of tremendous interest, both stratigraphically and botanically-phylogenetically"; Prof. S. N. Tyuremnov comments that it is of exceptional interest as a great contribution to paleobotany; Prof. L. I. Savich-Lyubitskaya believes it to be of great value to biologists and botanists; L. Sh. Davitashvili of the Georgian Academy of Sciences, evalutes it as a great scientific event; A. I. Ketova, a paleobotanist at Leningrad University, notes that this work has solved many problems in the morphology, anatomy, and systematics of plant remains hitherto unnoticed. In the opinion of Prof. B. B. Rodenford, I. F. Neyburg's conclusions on the climate of the Angarides are quite significant and in complete agreement with his own data obtained from the study of insects. Paleobotanist S. V. Sukhov draws attention to the new methods of studying plant remains used by M. F. Neyburg, and believes them to be of great importance in further studies in this field.

Prof. G. Erdtman, Stockholm, points to the great importance of this work and writes that it is being studied in his laboratory. Prof. O Hoeg of the University of Oslo notes the value of this work and comments on the excellency of its illustrations.

Prof. T. Harris, the University of Reading, England, stresses the value of the material studied and states that M. F. Neyburg has altered, with a single stroke, the prevailing paleobotanical concepts — a rare achievement for a scientist.

Li. Sin-Sue, a paleobotanist at the Academy of Sciences, Chinese People's Republic, comments on the great importance of this discovery of rare Bryophyta remains.

Paleobotanist B. Lundblad (Stockholm) expresses the hope that the Russian discoveries will stimulate the search for new Paleozoic mosses in other countries as well.

Prof. K. Megdefrad (Botanical Institute in Tübingen, GFR) regards the M. F. Neyburg studies as the most important achievement in paleobotany in recent years.

Editors

REPLY TO THE COMMENTS OF V. I. KITSUL AND M. A. BOGOMOLOV ON MY ARTICLE, "CONTACT-INFILTRATION SKARNS NEAR THE KONDER MASSIF CARBONATITE BODIES"[5]

by

G. V. Andreyev

In their review published in issue 1, 1961 of this journal, V. I. Kitsul and M. A. Bogomolov bring up a number of critical points refuting, in their opinion, some of my basic conclusions.

They note that "carbonate rocks rest exclusively on metamorphic rocks enclosing the massif", although they are familiar with the 1:25,000 geologic map compiled by A. N. Milto, A. A. Yemel'yanov, and G. V. Andreyev in 1957. This map shows the carbonate rocks occurring among koswites and exocontact gneisses, granitized sandstones, siltstones, and shales of the Omninsk formation. The 1958 work by G. V. Andreyev established the occurrence of carbonate rocks among the diorites and granite pegmatites. Instead of checking the data of earlier students, Kitsul and Bogomolov arbitrarily confine the carbonates to metamorphic rocks — gneisses and quartzites.

The fact is that the northwestern exocontact shows carbonate rocks traceable over considerable distances among the Ol'ninsk siltstones and shales, and at higher elevations than the gneisses. In addition, there are thin sections of koswites made from the bedrock on both sides of a carbonate vein exposed in the south part of the massif.

Carbonate veins in the southern and northwestern exocontacts contain isolated blocks of near-skarn rocks containing inclusions of koswites showing their typical sideronitic structure. The carbonate rocks contain up to 15% magnesium.

Variation diagrams of the columns consisting of carbonate rocks, skarns, and

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[5] Otbet na Zamechaniya V.I. Kitsula i M.A. Bogomolova po Stat' i "Kontaktovo-infil'tratsionnyye Skarny Vblizi Karbonatitovykh Tel Konderskogo Massiva", (pp. 112-113).

[ko]swites, illustrate the accumulation of silica [an]d alumina in the central zones. Petrographic [st]udies have shown that metasomatic processes [pr]oceed everywhere from carbonate rocks to [th]e enclosing rocks. This relationship does not [ho]ld true in the normal formation of mag[ne]sian skarns. Minerals of skarns and of near-[sk]arn rocks developed on koswites always show [th]e presence of up to 0.1% strontium, while it [is] absent in the koswites themselves. Stront[iu]m could have come from the carbonate rocks [on]ly — a process possible only in the pro[gr]essive replacement from these rocks to the [ko]swites.

[V.] Kitsul and Bogomolov utterly disregard the [pr]esence of fine-grained gneissoid rocks in the [ex]ocontact metamorphics. Petrographic studies [ha]ve established that these rocks were formed [du]ring the granitization of the Ol'ninsk sand[st]ones. The presence among the latter of un[alt]ered carbonate rocks, unconformably with [re]spect to the enclosing rocks, can be explain[ed] only by a magmatic origin for these car[bo]nates.

In their mineral composition, the Konder [ma]ssif carbonate rocks are not common [ma]rbles and calcifers as Kitsul and Bogomolov [as]sert. Perovskite, a typomorphic carbonatite [mi]neral, has been observed in them.

The high magnesium content, up to 25.5%, [is] a feature of the chemical composition of [th]ese carbonate rocks. No such figure has [be]en obtained for dolomitic marbles. It is to [be] remembered that the theoretical magnesium [co]ntent in dolomite is 21.7%. Higher contents [of] strontium (up to 0.3%) and niobium (up to [0.]0038%) have been observed in the Konder [ma]ssif carbonates. All these facts suggest a [ma]gmatic origin for the carbonate rocks.

It also should be noted that these carbonate [ro]cks are quite similar to carbonatites, in [th]eir mineral and chemical composition and [ce]rtain geophysical properties. Carbonate [ro]cks are known to be genetically associated [wi]th a central type massif of normal ultra[ba]sic rocks.

Now, the Konder massif is very much like [th]ose central type intrusions judging from the [st]ructure and composition of some of its rocks [(d]unites, koswites, apatite-biotite-titano[ma]gnetite-pyroxene rocks) and by its geologic [po]sition (margin of a platform).

For all of these reasons, the use of the [te]rm, "carbonatite", to describe the rocks of [th]e Konder massif — first made by V. V. [Ar]khangel'skiy, A. A. Yemel'yanov, and A. G. [Ga]ts, in 1956 — is justified.

The erroneous opinion of Kitsul and Bogo[m]olov on the nature of these carbonate rocks is based apparently on inadequate study. Inasmuch as the carbonate rocks are magmatic, while the replacement processes in metasomatic rocks near the carbonate veins go in the direction of the enclosing rocks (as confirmed by petrographic observations and by analysis of the variation diagrams compiled from metasomatic columns). It is easy to misinterpret the skarns as a product of carbonate solutions reacting of the enclosing rocks.

We disagree with the observation by Kitsula and Bogomolov on the behavior of Al_2O_3 in the metasomatic process. In our mobility series, silica is more inert than alumina, while the reverse is shown in D. S. Korzhikov's work. However, that author states that his mobility series should not be regarded as dogma and that it is subject to modification. Incidentally, an article by A. S. Pavlenko (this journal, No. 1, 1959) notes the greater inertia of silica as compared to alumina.

Thus the remarks of V.I. Kitsul and M. A. Bogomolov concerning the main premises of the article, "Contact-Infiltration Skarns Associated With the Carbonate Bodies in the Konder Massif", are without substance.

Received, 24 February 1961

LETTER TO THE EDITORS[6]

by

I. P. Kushnarev, and A. B. Kazhdan

In an article published in No. 9, 1959, of this journal, N. P. Vasil'kovskiy objects to our corrections and changes suggested for his stratigraphic scheme and unjustly reproaches us for noting only its contradictions. The truth is that our purpose was not to review his monograph but merely to present the results of our long years of field and office work on the stratigraphy of that region. What is more, in regarding N. P. Vasil'kovskiy's work as an outstanding contribution to the Upper Paleozoic stratigraphy and volcanology of the southwestern spurs of the Northern Tien-Shan, we have acknowledged thereby its great merit.

A comparison of our and N. P. Vasil'kovskiy's schemes shows the following principal differences: 1) we regard as one, the three first formations (Arkutsay, Uya, and Myn-Bulak); 2) we see no reason for replacing the Sarysiyum formation by the Nadak or the Kushaynak; 3)

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[6]Pis'mo v Redaktsiyu.

we do not recognize the Ravash formation as independent. The remaining differences between our and the preceding N. P. Vasil'kovskiy schemes [1, 2] have been eliminated by the author himself [3].

1. N. P. Vasil'kovskiy has not presented any evidence supporting his assertion to the effect that the Uya and Myn-Bulak formations are non-contemporaneous [3]; he is also wrong in stating that our only evidence for their correlation is the similarity in their rocks. We have always emphasized the necessity of a careful study of changes in facies and thickness for each member of a formation. It is exactly the omission of these two factors that has led N. P. Vasil'kovskiy to the erroneous conclusion [3] that a 1000 m thick interval (mostly sedimentary) at the base of the Uya formation, in the lower Ugam course, is missing because of erosion. As the matter of fact, what we have here is another instance of the change from the sedimentary Dzhigergen facies of the Uya formation — south of the Dzhigergen Saya formation — to the effusive-sedimentary series of the same formation — a change he, himself, noted [1]. A. B. Kazhdan demonstrated in 1955 that this phenomenon also takes place farther south, i. e., toward the mouth of the Ugam, and leads to almost the same change from sedimentary to effusive rocks. N. P. Vasil'kovskiy also insists on assigning to the Arkutsay porphyrites the rank of an independent, locally developed formation, while it is, in fact, a facies of the lower Uya formation, as demonstrated earlier by A. B. Kazhdan.

The fact that rocks of the same age underlie both the Uya and Myn-Bulsk formations (Visean — base of the Namurian) led the Conference for a Unified Stratigraphic Scheme of Central Asia, to place these two formations side-by-side, rather than consecutively, as N. P. Vasil'kovskiy does.

He misleads the reader by stating that the upper age limit of the Uya formation is the Bashkirian stage, with the Moscow stage of the Carboniferous as the upper limit of the Myn-Bulak formation. As a matter of fact, there are no faunally characterized sediments of that age in the upper, largely effusive, 1200-1300 m thick interval of both formations, let alone the formations unconformably on them. Resting unconformably on the two is the Akchinsk effusive formation containing occasional plant remains which define its age as Middle to Upper Carboniferous.

Considering that the Uya and Myn-Bulak formations have never been observed in the same section, and also considering the data presented, one cannot but conclude that we deal here with two facies of the same formation, formerly designated by different names.

2. N. P. Vasil'kovskiy denies the independence of the Sarysiyun formation, believing it to be the basal beds of the Oyasy formation. His argument is the absence of intrusions cutting the first formation but not the second. Now, inasmuch as its type section shows clean-cut, angular unconformities of 10-45° between the two, the Sarysiyum formation rocks can not be the basal beds of the Oyasay formation. Elsewhere, the latter rests on most diversified deposits down to the Silurian. Furthermore, many geologists have established that the Gushsay granodiorite-porphyries and Dukent stocks of granite-porphyrites and granosyenite-porphyries in the basins of the Dukent and Maylikatan Rivers, cut the Sarysiyun formation (as well as Nadak conglomerates in the Kuramin range), and are overlain by the Oyasay formation, whose base contains them in pebbles and angular fragments. These facts are positive proof of the independence of the Sarysiyun formation.

We have shown [4] that the Nadak conglomerates cannot be assigned to the Ravash formation since they are correlative with the Sarysiyun deposits in composition and position in the stratigraphic column. We see no reason, therefore, for replacing the regionally developed Sarysiyun formation by the so-called Nadak formation — in effect a mere substitution of one name for another. Nor can the Sarysiyun formation be correlated with the Kushaynak sequence of Eastern Karamazor, since its conformity with other sections was correctly assigned to the Akchin formation, by Ye. D. Karpova. Corresponding to the Sarysiyun formation in this region is the Kysyltau formation of Ye. D. Karpova, unconformably on various intervals of the Akchin formations and overlain unconformably by the Oyasay formation. N. P. Vasil'kovskiy [3] is utterly without justification when he lowers the position of the Sarysiyun formation from C_3 to C_2—C_3, and of the Akchin formation from C_2—C_3 to C_2.

3. N. P. Vasil'kovskiy does not present any new arguments for an independent Ravash formation. In speaking of the Ten'ga village area (lower course of the Gava-say), he refers to the long well-known work of A. S. Makarov, N. P. Podkopayev, and Z. P. Artemova, who correctly state that there are no azimuthal unconformities between rocks assigned to the Shurabsay and Ravash formations and that the few dips measured in isolated areas differ by 35-40°. However, N. P. Vasil'kovskiy fails to mention that I. P. Kushnarev, in 1952, and A. F. Utkin, in 1954, established here the presence of a gradual steepening of the dip, which is a common occurrence on the limb of a fold.

The Kassan graben area was mapped in detail by A. F. Utkin in 1958 and was also visited by I. P. Kushnarev. Particular attention was paid to the relationship between the several Permian units. Again, no unconformities have

ɔn observed which would warrant a differ- iation of the two formations — Shurabsay d Ravash.

We omit the remaining minor points — like- se unsupported by field data. What has been d is enough to show the inconsistency of P. Vasil'kovskiy's arguments for his basic emises and the lack of justification in his ticism of ours.

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Received, 28 December 1960

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The following changes have been recommended by the author for the translation of the Russian version of his paper: "Phase Relationship in Peridotites of Dawros (Ireland) and Belhelvie (Scotland)", by A. T. V. Rothstein. Izvestiya Akad. Nauk SSSR, Seriya Geologicheskaya, 1961, No. 3.

Page	Column	Line from top	Line from bottom	Printed	Correction
52	2	1	—	with inclusions	including
53	Fig. 2	—	8	extrusive	—
57	2	7	—	considering	if it is considered
57	2	—	5	blends	mixtures
58	1	—	9	toward	through
58	2	—	1	borderline	boundary
59	2	6	—	basalt	basal
59	2	21	—	occurrence for	meeting of
59	2	—	5	90°	900°
61	1	6	—	deterioration	failure
61	1	—	19	the reaction	into reaction
61	2	—	4	leists	laths
61	2	—	3	ratio	relationship
62	1	footnote		this embraces	this case embraces
62	2	—	7	such thing as a	complete
63	1	2	—	enstatite and pyroxene	enstatite and diopside
63	Fig. 17b	2	—	bronzite and primary diopside	primary bronzite and diopside
63	1	8	—	solvus	solidus
63	1	13	—	of	to
63	1	—	3	recrystallized	crystallized
63	2	6	—	Ram	Rhum